MINERALOGICAL MAGAZINE

JOURNAL OF THE MINERALOGICAL SOCIETY

Vol. 32 September 1961 No. 254

Some new platinoid-rich minerals, identified with the electron microanalyser.

(With Plate XVII.)

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[Read 8 June 1961.]

Summary. Minerals approximating to the formulae $PtSb_2$, PtSb, Pt(Sb,Bi), $(Pt,Ir)As_2$, $Pt(Ir,Os)_2As_4$, Pd_2CuSb, Pd(Sb,Bi), Pd_8CuSb_3, $Pt_4Sn_3Cu_4$, and $(Fe,Ni)_2S$ have been discovered as fine intergrowths in platinum concentrates from the Driekop mine, Transvaal, South Africa. Conventional ore microscopy also proved the presence of most of the known minerals of the platinum paragenesis. The name *geversite* is proposed for the phase $PtSb_2$.

MICROSCOPIC investigation of platinum concentrates from the Driekop mine, Transvaal, South Africa, which the author collected during a visit to the mine in 1957, revealed the presence of a number of unknown minerals as well as a wide range of sulphides and other established members of the platinum paragenesis. The unknown minerals occur in very fine intergrowths, which made the separation of individual phases for analytical purposes or X-ray work impossible. Thus, it was not until an X-ray microanalyser was put at the author's disposal that the chemical composition of the minerals could be determined. Based on this work, the results of which frequently fitted the indications given by ore-microscopic evidence, it was possible to establish the formulae of ten new minerals. Only a few synthetic products corresponding in composition to these minerals are known.

Details of the microanalyser used, which was designed by Castaing and constructed by the CAMECA Company, Courbevoie (Seine), France, are to be found in Castaing (1960). The principles and application of the microanalyser are now becoming so well known (Guillemin and Capitant,

1960; Capitant and Weinryb, 1960), that only a few of the main points will be reviewed here. The atoms of a sample (in this case polished sections of ore concentrates embedded in 'Araldite' resin) are excited by an electron beam focused on an area of about one square micron. Returning to their stable state, they generate X-rays, the spectra of which are analysed in two curved-crystal spectrographs. One of these, for elements with an atomic number of 11 to 17 inclusive, uses a mica strip and a gas flow proportional counter. An alternative spectrograph, for analysing elements with atomic numbers above 17, uses a quartz strip and a Geiger-Müller counter. Elements with atomic numbers below 11 cannot be detected with the equipment in question. For quantitative analysis, the intensity of each spectral peak is measured and compared with readings obtained from pure element standards, and as the sample holder is fitted with a reference sample carrier that takes up to 40 standards, the measurements can be carried out with considerable speed. Corrections of the recorded values have to be made in some cases because of the differing mass absorption coefficients of different elements. A procedure for this purpose has been developed by Castaing and the corrections increase only slightly the time needed for analysis. The accuracy of the quantitative measurements is ± 5 %. In those cases where the sum of the percentages of elements determined does not approximate 100 %, some elements may have remained undetected during analysis. This applies especially to elements with atomic numbers below 17 (i.e. elements lighter than Cl). In those instances the formulae calculated must be considered as only approximate.

Geological environment. The Driekop platinum mine is situated in the norites of the critical zone of the Bushveld complex in eastern Transvaal (see sketch-map, fig. 1). The host rock of the platinum minerals is a pipe-like body of dunite that dips about 80° NE.—roughly normal to the pseudostratification of the norites. On the surface, the dunite pipe stands out as a small hill 100 ft. high above the surrounding norite country. Only a small portion of the dunite is to be considered as workable ore and the cross-section through the pay-zone measures about 60×80 ft. It is remarkable that the occurrence of platinum pipes (Driekop, Onverwacht, Mooihoek) in this part of the Bushveld complex is confined to a straight line about 10 miles in length and striking N. 35° W. Wagner (1929), who dealt in detail with the geological and genetic implications, suggested relations to a potential line of weakness in the underlying rocks.

The Driekop pipe differs from the other two mentioned above in that

it shows neither a concentric structure nor differentiation of the dunite into iron-rich and iron-poor fractions. Heckroodt (1959), dealing with the petrography of the rocks in the Driekop area, failed to establish any relationship between the platinum content of the dunite and the com-

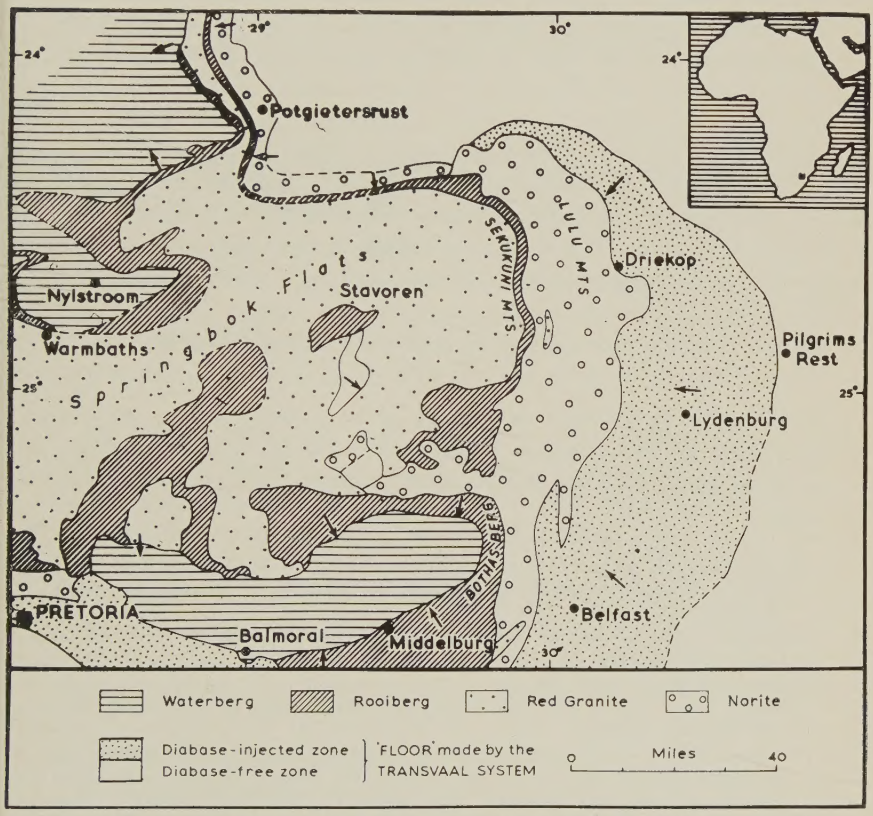


Fig. 1. Geological sketch-map of the eastern part of the Bushveld igneous complex (modified after du Toit).

position of the olivine. According to Heckroodt (1958) the composition of the olivine in the ore-body varies from Fo_{71} to Fo_{84}, but does not display any concentric distribution.

Rich parts of the ore-body carry up to 10 p.p.m. platinum. It is thus evident that an efficient investigation of the platinum minerals cannot be undertaken on dunite specimens, but only on mine concentrates.

The ore paragenesis. As the main purpose of this paper is to draw

attention to the new platinoid minerals, the known and easily recognized minerals are but briefly mentioned. The concentrates investigated proved to contain a wide variety of ore minerals. Amongst the sulphides, chalcopyrite and pyrrhotine predominate, while among the platinoid minerals, sperrylite (often in euhedral crystals), $PtSb_2$ (which has not been known before), and native platinum are the most important. Quantitative analysis with the electron microprobe showed that the

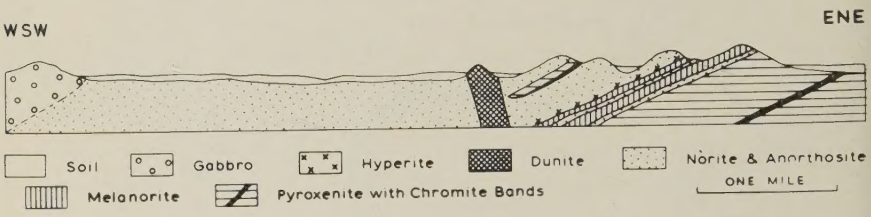


Fig. 2. Profile through the Driekop area (after Heckroodt).

native platinum contains an average of 18 % Fe, and may thus be designated as ferroplatinum. Associated with the above-mentioned principal constituents are the following minerals, mostly in small quantities:

Pentlandite, (Fe,Ni)S.

Valleriite, $Cu_3Fe_4S_7$ (frequently as lamellae and as grains of considerable size in chalcopyrite).

Covelline, CuS (obviously an alteration product of chalcopyrite).

Braggite, (Pt,Pd,Ni)S.

Cooperite, PtS (pl. XVII, fig. 4).

Laurite, RuS_2. Rather rare, but easily recognized because of its hardness and difficulty in polishing.

Stibiopalladinite, Pd_3Sb, which Ramdohr (1960) considered the most widespread platinoid mineral; this is not confirmed by the present investigation. Some of the new minerals described later, especially PtSb and Pt(Sb,Bi), may frequently have been wrongly identified as stibiopalladinite.

Osmiridium. Extremely hard, silvery-white lamellae, mostly of comparatively large size, in grains of native platinum.

Iridosmine, variety osmite. Easily distinguished from osmiridium because of its distinctly reddish anisotropy colours and a touch of bluish-grey tint in plane polarized light (pl. XVII, fig. 4). The ore-microscopic characteristics of osmite have for the first time been observed by Picot (1961, personal communication), using material the composition of which has been determined by the electron microanalyser.

Native copper. Large independent grains as well as small inclusions in native platinum and sperrylite.

Native gold. Small flakes in some of the platinoid minerals.

Chromite. Widespread.

THE NEW MINERALS

No less than ten new minerals were found in the concentrates. Except for a very interesting sub-sulphide of iron and nickel all are compounds of the platinum metals with As, Sb, and Bi.

Minerals with Pt and Sb as main constituents.

Geversite, $PtSb_2$. Reflectivity is lower than that of native platinum, but higher than that of sperrylite. The colour is light grey (pl. XVII, fig. 3). No pleochroism and no anisotropy effects are to be observed; this indicates isotropic character and hence cubic symmetry. The hardness is similar to that of native Pt, perhaps slightly higher.

The microprobe analysis gave Pt 45·0, Sb 51·5 %, corresponding with the formula $PtSb_2$. Fig. 4 shows the X-ray spectrogram of $PtSb_2$; obviously this is the antimony equivalent of sperrylite, $PtAs_2$. The most distinct differences from sperrylite—lower hardness and slightly higher reflectivity—are the results of replacement of As by Sb in the lattice.

The system Pt–Sb has been investigated by Friedrich and Leroux (1909) and by Nemilov and Voronov (1935, 1936); the main results are summarized in Hansen (1958) (see fig. 3). Synthetic $PtSb_2$ (Thomassen, 1929*a*) is known to crystallize, as sperrylite does, with a pyrite-type lattice, has a 6·428 Å. and a melting-point of 1 230° C., and contains 44·47 % Pt. In polished section it looks grey and quite homogeneous, is isotropic, and has a hardness of $4\frac{1}{2}$ to 5. As a microprobe analysis of sperrylite shows, small amounts of Sb can replace As; in pl. XVII, fig. 6, $PtAs_2$ containing 2·8 % Sb is to be seen. On the other hand, Sb in $PtSb_2$ can be replaced by Bi, resulting in a slight increase in reflectivity; pl. XVII, fig. 2, shows $PtSb_2$ with 7·2 % Bi.

$PtSb_2$ is widespread, occurring in small isolated grains as well as intergrown with other members of the paragenesis. Its widespread association with other Pt–Sb minerals, as shown in pl. XVII, fig. 3, is noteworthy. In certain cases, PtSb and Pt(Sb,Bi) can be observed as very fine myrmekitic inclusions in $PtSb_2$. Both these features underline the close genetic relationships of the Sb-rich members of the platinum paragenesis. $PtSb_2$ is generally the earliest of the Pt–Sb minerals, followed by PtSb and Pt(Sb,Bi). This may imply a significant correlation with the melting-points of the minerals which are 1 226° C. for $PtSb_2$, ~ 1 050° C. for PtSb and probably lower for Pt(Sb,Bi) and which probably indicate an order of crystallization from a magma. $PtSb_2$ has

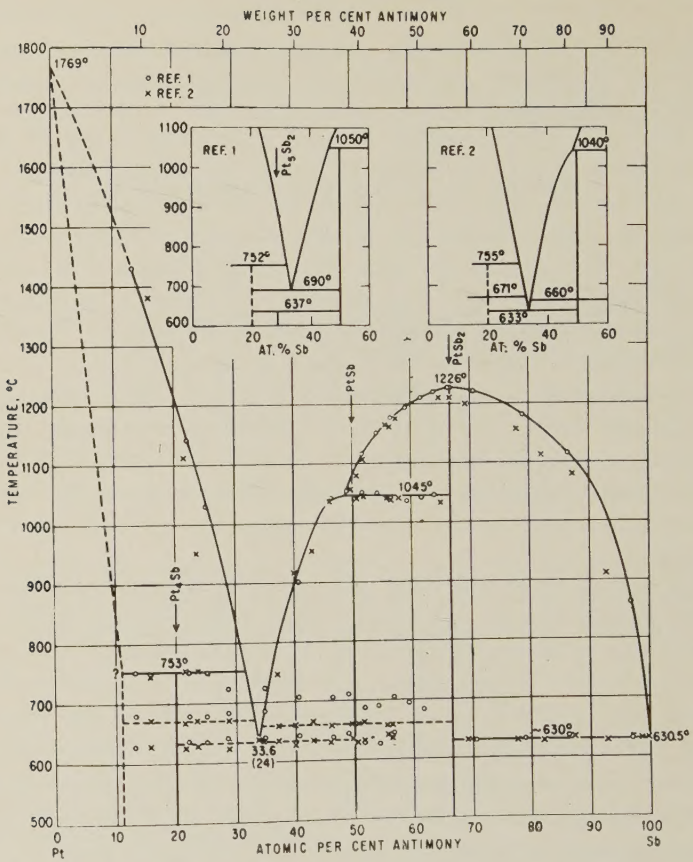


FIG. 3. The system Pt–Sb. Ref. 1 refers to Friedrich and Leroux, 1909, and ref. 2 to Nemilov and Voronov, 1935, 1936. (By permission from M. Hansen, 1958, Constitution of binary alloys (McGraw-Hill Book Co.).)

not previously been known to occur as a mineral, and the name geversite (pronounced GĒ·VERZAIT) is proposed for it, in honour of the eminent South African geologist, Prof. T. W. Gevers.

Pt(Sb,Bi). This mineral has a reflectivity lower than that of native platinum, but, as will be evident from pl. XVII, fig. 3, higher than that of $PtSb_2$. The colour has a distinct pinkish tint. Pleochroism is to be observed easily in air, the colour changing from pink to light brown (in the latter position it is indistinguishable from PtSb). Under crossed nicols, intense anisotropy effects, with pink colours, are to be observed.

In oil immersion both pleochroism and anisotropy are much more pronounced. The hardness is slightly less than that of native platinum. This mineral, especially in very fine grains, may frequently have been mistaken for stibiopalladinite. The intense anisotropy suggests a layer lattice probably of hexagonal symmetry. The fact that both PtBi and PtSb crystallize with nickel-arsenide lattice supports this view. The microprobe analysis gave the following values: Pt 50·5, Sb 26·4,

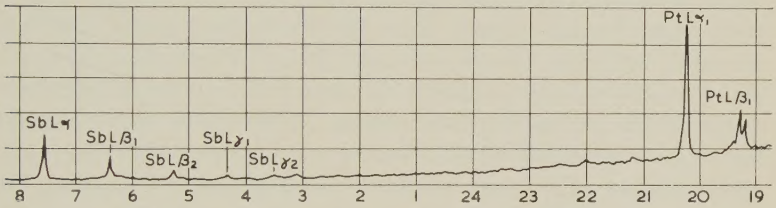


FIG. 4. X-ray spectrogram of $PtSb_2$ (geversite).

Bi 15·3 %. This suggests the formula $Pt_7Bi_2Sb_6$, or, simplified, approximately Pt(Sb,Bi). After geversite, this mineral is the most widespread of the Pt–Sb group. It does not, however, occur as independent grains, but mostly associated with $PtSb_2$ and PtSb.

PtSb. The reflectivity of this mineral is lower than that of Pt(Sb,Bi), but still much higher than that of pyrrhotine (for comparison, see pl. XVII, fig. 3). The colour is light brownish with a touch of pink. Pleochroism is absent. Under crossed nicols, anisotropy effects are to be observed, the intensity of which, however, is lower than in Pt(Sb,Bi) or pyrrhotine. The hardness is lower than that of native platinum. Quantitative analysis showed Pt 50·5, Sb 34·7 %. The most likely formula is PtSb. The results of microscopic examination of the mineral and the chemical data are in good agreement. Thomassen (1929*b*) made synthetic PtSb[1] and described it as a 'yellow mass' in plane polarized reflected light. He found *a* 4·130 Å., *c* 5·472 Å., *c*/*a* 1·325 (nickel arsenide type lattice).

Minerals containing Pt and (Ir,Os).

(Pt,Ir)As$_2$. The colour is distinctly greyish, the reflectivity high, but lower than that of $PtSb_2$ (pl. XVII, fig. 2). There is no pleochroism and anisotropy effects are absent. The mineral obviously has cubic symmetry.

[1] It is interesting to note that synthetic PtSb has recently been found to become superconducting at ~2·1° K. (Matthias, 1953*a*, 1953*b*). The same applies to the related NiAs-type lattice compounds PtBi (~1·21° K.), PdSb (~1·5° K.), and PdBi (3·74° K.).

Its hardness is higher than that of Pt and $PtSb_2$, probably around 6 in Mohs' scale. The analysis gave the following results: Pt 47·0, As 46·8, Ir 5·0 %. It is possible that this mineral should be considered as an iridian sperrylite, though Ir has not hitherto been recorded in sperrylite. The analysis results can be fitted approximately into the $PtAs_2$ formula, and a certain degree of replacement of Pt by Ir might well be possible.

$Pt(Ir,Os)_2As_4$. The typical feature of this mineral is a distinct touch of brownish colour in reflected light. Its reflectivity is similar to that of sperrylite, and lower than that of Ir-rich $PtAs_2$. The hardness is higher than that of Ir-rich $PtAs_2$. Increasing amounts of Ir and Os are obviously responsible for the increase in hardness. Examination of the mineral in both plane polarized light and under crossed nicols did not reveal any signs of anisotropy, again indicating cubic symmetry. Lamellae of native osmium (iridosmine, var. osmite) are not infrequently to be observed within this mineral (pl. XVII, fig. 1). It is difficult to decide whether these are due to decomposition or to unmixing processes.

The analysis gave Pt 22·0, As 32·5 %, and considerable amounts of iridium and osmium are also present. A quantitative determination of the percentages of Ir and Os was not possible as the microprobe analyser revealed remarkable changes in the content of these two elements within a few microns distance. The Ir and Os values recorded along a traverse from A to B are to be seen in fig. 6. The complete mutual replaceability of these two elements is very well demonstrated by high Ir values corresponding to low Os values, and vice versa. The observable optical properties of the minerals are not affected by these changes in chemical composition, as is evident from pl. XVII, fig. 2. The great similarity in atomic radius and chemical properties of these two elements makes it understandable that variations in their relative proportions would not appreciably influence the physical properties of the mineral concerned. Nevertheless, if one considers Ir and Os as the only probable constituents besides Pt and As, then $Pt(Ir,Os)_2As_4$ becomes likely as a formula.

Generally these Ir- and Os-rich minerals are by no means as widespread in the concentrates as, for instance, the Pt–Sb compounds. They occur mostly together in anhedral grains. Frequently $Pt(Ir,Os)_2As_4$ is to be observed as lamellae in native platinum as well (pl. XVII, fig. 1).

Mineral with Pt, Cu, and Sn.

This mineral is extremely rare and is difficult to recognize under the microscope. Its reflectivity is very high, but still distinctly lower than

that of native Pt, and slightly lower than that of Bi-rich $PtSb_2$. It has a touch of pink, similar to PtSb. Under crossed nicols, in oil immersion, anisotropy without special colour effects is visible.

The hardness is lower than that of $PtSb_2$ and not very different from that of native platinum. Quantitative analysis yielded unexpected results: Pt 51·7, Sn 22·0, Cu 16·8 %. The only natural association of the two elements Pt and Sn that has hitherto been recorded is in niggliite, Pt_2Sn_3. Pt–Sn alloys with varying Pt–Sn ratios are reported in Mellor (1937) as having been prepared by A. F. Gehlen, and in Thorpe (1950), viz. Pt_3Sn_8, Pt_2Sn_3, PtSn, Pt_3Sn. The formula calculated from the above values is $Pt_4Sn_3Cu_4$. It is interesting to note the occurrence of a second mineral containing Sn at the 'basic end' of the differentiation sequence. The presence of small amounts of Sn in platinum metal concentrates from the Sudbury ores has been proved by chemical analysis (personal communication, Dr. B. Taylor, Acton Platinum Metals Refinery of the International Nickel Co., Ltd.). It may seem remarkable that tin has not previously been noted in the rocks concerned but it must be realized that $Pt_4Sn_3Cu_4$ is very rare even in relation to the other platinoid minerals. The total platinoid mineral content is, as previously noted, only about 10 p.p.m.

Palladium minerals.

Pd_2CuSb. Characteristic features are its strong violet colour and a reflectivity much lower than all the other minerals of the paragenetic association (pl. XVII, fig. 5). They are, as usual, more pronounced in oil immersion. Under crossed nicols, good anisotropy effects with orange colours are to be observed. The hardness is not very different from that of native platinum. The analysis gave the following results: Pd 53·5, Sb 31·3, Cu 16·0 %. This corresponds with the formula Pd_2CuSb. Any close similarity with stibiopalladinite is ruled out by both optical and chemical data. Natural substances of this composition have not so far been recorded. It is one of the rarest constituents in the concentrates, always occurring in intergrowths with other platinum minerals.

Pd_8CuSb_3. Intergrown or closely associated with the above mineral there occur grains and lamellae of higher reflectivity and a pale yellowish colour (pl. XVII, fig. 5). Anisotropy effects are well pronounced, especially in oil, but without recognizable colour. The reflectivity is still slightly lower than that of geversite, as may be seen in pl. XVII, fig. 5. The hardness is similar to that of native platinum. Quantitative

micro-probe analysis gave the following results: Pd 62·0, Sb 30·0, Pt 6·0, Cu 5·0 %. These values correspond with the formula Pd_8CuSb_3. Small amounts of Pd may be replaced by Pt. These two palladium minerals are obviously closely related, as indicated by the similarities in chemical composition as well as by their frequent mutual intergrowth.

Pd(Sb,Bi): a light yellowish mineral with high reflectivity and not showing so pale a colour as Pd_8CuSb_3 does. No pleochroism, but anisotropy effects of medium intensity with gold-yellow colours under crossed nicols. Hardness is distinctly lower than $PtAs_2$ (pl. XVII, fig. 6). The microprobe analysis gave the following results: Pd 43·6, Sb 25·2, Bi 32·2 %, suggesting the formula $Pd_8Sb_4Bi_3$. Simplified, this formula could be given approximately as Pd(Sb,Bi) with either an excess in Pd or a deficiency in (Sb,Bi). Accordingly, this mineral might be considered as the palladium equivalent of Pt(Sb,Bi), probably crystallizing with a hexagonal lattice. Thomassen (1928) reported synthetic PdSb as crystallizing with NiAs-type lattice (*a* 4·070 Å., *c* 5·582 Å., D 9·4); PdBi, according to Zhuravlev (1957), is orthorhombic with *a* 7·203, *b* 8·707, *c* 10·662 Å.

Pd(Sb,Bi) does not occur as independent grains, but only intergrown with Pt-arsenides and other Pt-minerals, sometimes in complex textures, (pl. XVII, fig. 6).

Nickel-iron subsulphide.

$(Fe,Ni)_2S$. The occurrence in small grains, intergrown with other members of the paragenesis, of a mineral resembling pyrrhotine in colour is worthy of note. Its reflectivity, however, is slightly higher than that of pyrrhotine (pl. XVII, fig. 2). Under crossed nicols, anisotropy effects without colours are observed. The hardness is comparatively low. The chemical composition is Fe 51·0, Co 1·0, Ni 28·0, S 22·5 %, corresponding with the formula $(Fe,Ni)_2S$. A section of the X-ray spectrogram is shown in fig. 5, indicating only the metallic elements. It must be considered whether small amounts of sulphur available in the magma were responsible for the formation of this unusual mineral. Mellor (1936) mentions the synthetic preparation of nickel hemisulphide, Ni_2S, by heating nickel sulphate in a current of hydrogen. He further states, following Rose, that 'the pale yellow or brass yellow mass is not magnetic'. Similarly strong reducing conditions have to be postulated for the formation of $(Fe,Ni)_2S$. The presence of free hydrogen in ultrabasic rocks has been proved, for instance by Hahn-Weinheimer (1961). Finely dispersed native platinum might act as catalyst.

Ramdohr (1949) described a paragenesis of nickel sulphides at Trial Harbour, Tasmania. He found increasing amounts of sulphur present with decreasing temperature, covering the range from awaruite (sulphur:metal = 0) to pentlandite (sulphur:metal = 8:9). The possibility of $(Fe,Ni)_2S$ being an intermediate stage in this range has to be considered.

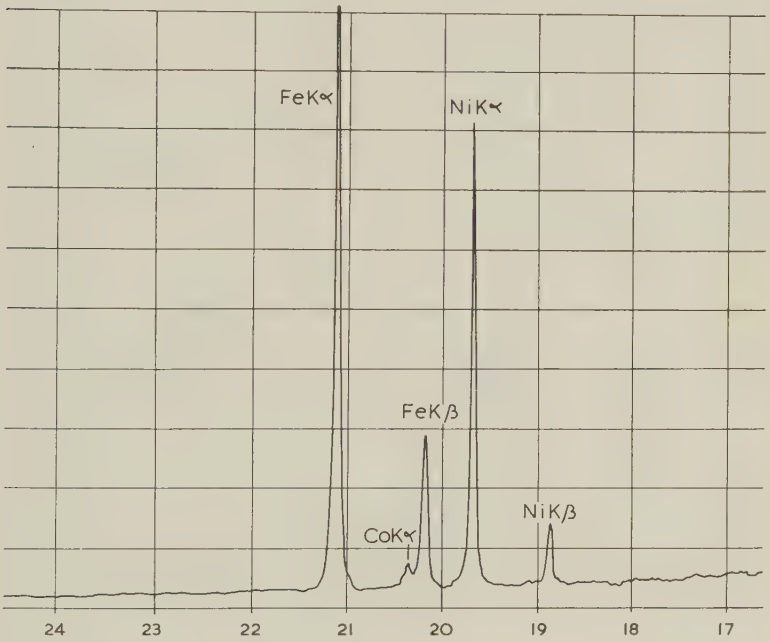


Fig. 5. X-ray spectrogram of $(Fe,Ni)_2S$.

Pl. XVII, fig. 2, shows $(Fe,Ni)_2S$ to be one of the earliest crystallized minerals, being surrounded by the whole sequence of later-formed phases.

Conclusions.

There can be no doubt that the substances described are new minerals. The author feels, however, that with the exception of geversite they should not be named until further investigations, which are being continued, have made it possible to separate amounts sufficient for X-ray analysis and the calculation of cell dimensions. Moreover, comparison with synthetically prepared substances might provide further data. The mineralogical description raised many problems, such as parageneticaland

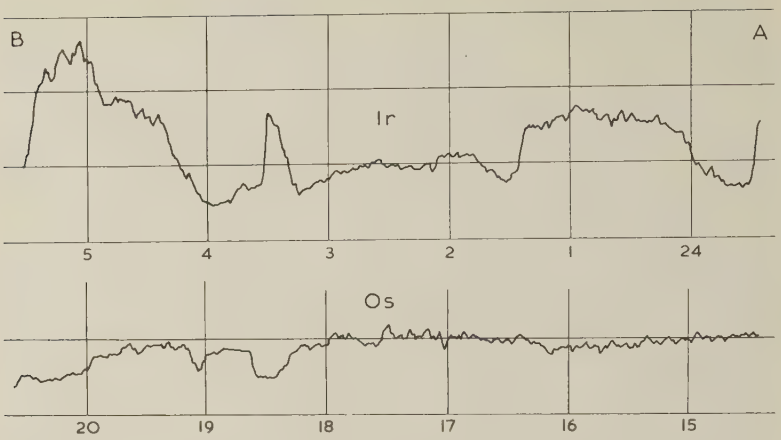


Fig. 6. The variation in Ir and Os concentrations within a single grain of $Pt(Ir,Os)_2As_4$; the traverse is from A to B in the grain marked in pl. XVII, fig. 2.

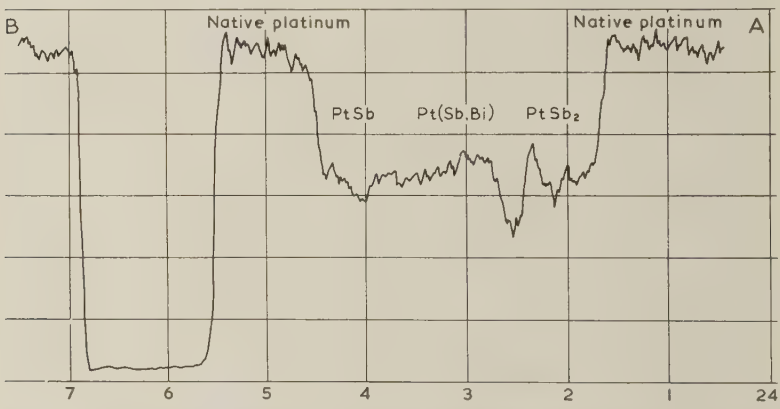


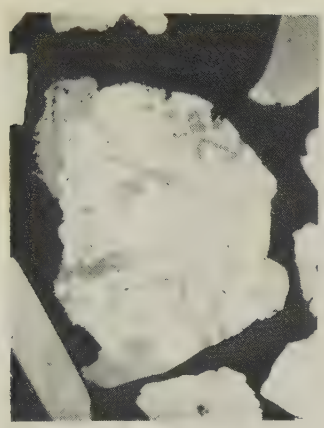
Fig. 7. The variation in platinum concentration along a traverse from A to B on the section shown in pl. XVII, fig. 3.

Explanation of Plate XVII.

Fig. 1. Lamellae of $Pt(Ir,Os)_2As_4$, medium grey, in native Pt (white). Native osmium (osmite) light grey, in the $Pt(Ir,Os)_2As_4$. Oil immersion, ×200.

Fig. 2. $(Fe,Ni)_2S$ (1); native Pt (2); $PtSb_2$ containing 7·2 % Bi (3); $(Pt,Ir)As_2$ (4); $Pt(Ir,Os)_2As_4$ (5). Oil immersion, ×360. The very small differences in reflectivity between the different minerals have been enhanced by special printing. Text-fig. 6 (*opposite*) shows the variation in Ir and Os from A to B, within a grain of $Pt(Ir,Os)_2As_4$.

Fig. 3. Native platinum, white; $PtSb_2$, Pt(Sb,Bi), lightest grey, PtSb; corre-



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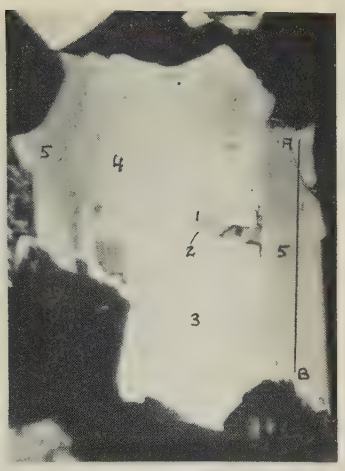

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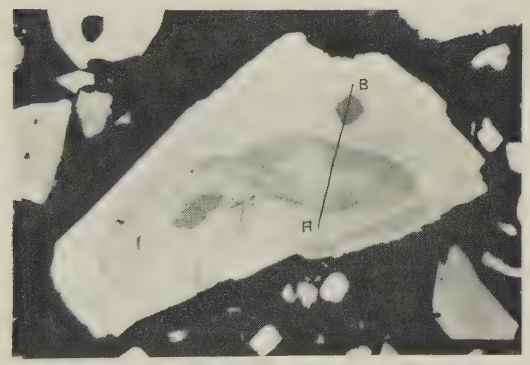

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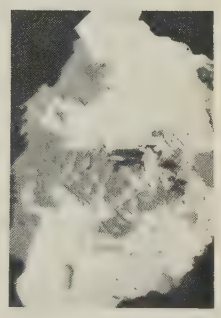

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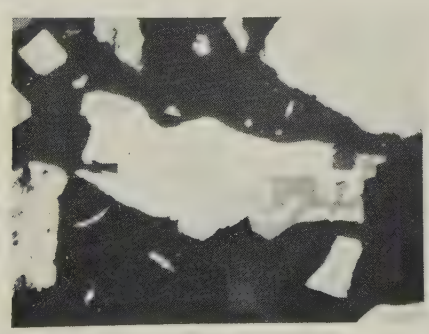

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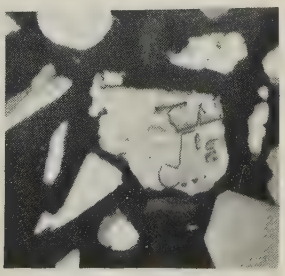

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E. F. Stumpfl: New platinoid minerals

age relations, which are deferred for the time being. The comparatively simple constitution of most of the new minerals described is remarkable. There is perfect agreement with some of the data available on Pt-Sb alloys. For future work on those minerals, however, that cannot be fitted into binary systems, greater difficulties are envisaged.

Acknowledgements. I am deeply indebted to Dr. C. Guillemin, in charge of the mineralogical section of the Bureau de Recherches géologiques et minières, Paris, for putting their electron microprobe at my disposal. Without Dr. Guillemin's help and generous hospitality, and the enthusiastic co-operation of MM. Phan, Picot, Bahezre, and Giraud, these investigations could not have been brought to a successful conclusion. Monsieur Bahezre kindly carried out the correction calculations for the microprobe analyses. I am grateful to Prof. P. Ramdohr, who, when I was working under him at Heidelberg, arranged for me to accompany him on his journey to South Africa in 1957 and made it possible for me to spend a few extra days at the Driekop platinum mine, where I appreciated very much the help of Mr. G. Onken, mine manager. My thanks are due to Prof. S. E. Hollingworth, Dr. E. A. Vincent, and Dr. J. R. Butler, who critically read the manuscript, and to the Direction Générale, bureau de Récherches géologiques et minières, Paris, for granting permission to publish those results that were obtained in their laboratories.

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sponding to traverse in text-fig. 7 opposite. $CuFeS_2$, medium grey grain left of centre; FeS, dark grey grain near B. Oil immersion, ×220.

Fig. 4. Native platinum, white; sperrylite, light grey; cooperite, dark grey. Oil immersion, ×460.

Fig. 5. Native Pt, white; $PtSb_2$, lightest grey; Pd_8CuSb_3, light grey; Pd_2CuSb, medium grey. Oil immersion, ×360.

Fig. 6. Pd(Sb,Bi), medium grey; $PtAs_2$, light grey, contains 2·8 % Sb. In centre small grain containing Ir 10, Pt 18, As 50 %. ×100.

TABLE I. Chemical and optical data for the new platinoid minerals and for $(Fe,Ni)_2S$.

	$PtSb_2$	Pt(Sb,Bi)	PtSb	$Pt(Ir,Os)_2As_4$	$Pt_4Sn_3Cu_4$	Pd_2CuSb	Pd_3CuSb_3	Pd(Sb,Bi)	$(Fe,Ni)_2S$
	%	%	%	%	%	%	%	%	%
Pt	45·0	50·5	50·5	22·0	51·7	—	6·0	—	—
Pd	—	—	—	—	—	53·5	62·0	43·6	—
Os	—	—	—	variable	—	—	—	—	—
Ir	—	—	—	,,	—	—	—	—	—
Fe	—	—	—	—	—	—	—	—	51·0
Co	—	—	—	—	—	—	—	—	1·0
Ni	—	—	—	—	—	—	—	—	28·0
As	—	—	—	32·5	—	—	—	—	—
Sb	51·5	26·4	34·7	—	—	31·3	30·0	25·2	—
Bi	—	15·3	—	—	—	—	—	32·2	—
Sn	—	—	—	—	22·0	—	—	—	—
Cu	—	—	—	—	16·8	16·0	5·0	—	—
S	—	—	—	—	—	—	—	—	22·5
Hardness	$\geqslant$ Pt	$<$ Pt	$<$ Pt	$>$ $PtAs_2$	$\sim$ Pt $<$ $PtSb_2$	$\sim$ Pt	$\sim$ Pt	$\sim$ $PtSb_2$	$<$ Pt
Colour (oil imm.)	white-greyish	light pink	light brown with pink tint	gray with brownish tint	white with pink tint	violet	pale yellow	light yellow	similar to FeS
Reflectivity	very high but lower than Pt	$<$ Pt $>$ $PtSb_2$	$\geqslant$ $PtAs_4$	$\leqslant$ $PtAs_2$	very high but lower than Pt	$\leqslant$ FeS	$<$ $PtSb_2$	high but lower than Pt	$>$ FeS $<$ $PtAs_2$
Pleochroism (oil imm.)	—	intense, pink–light brown	—	—	—	—	—	—	—
Anisotropy (oil imm.)	—	intense	visible	—	weak	distinct with orange colours	distinct without colours	distinct gold-yellow colours	weak
Crystal system	cubic	? hex.	hex.	cubic	?	? hex.	?	?	?

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Addendum: The author's attention has just been directed to three unnamed new minerals described by I. B. Borovsky, A. N. Deev, and I. D. Marchukova ([И.Б. Боровский, А. Н. Деев, и И. Д. Марчукова, Геол. Рудн. Месторожд., 1959, no. 6, p. 28]; abstr. in Amer. Min., 1961, vol. 46, p. 446) and by A. D. Genkin (А. Д. Генкин, ibid., p. 74). None of these minerals approximates in composition to any of those described above.

Petrology of the Scourie dyke, Sutherland.

(With Plate XVIII.)

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[Read 8 June 1961.]

Summary. Chemical data are given for four rocks from the Scourie dyke, Sutherland, and for the constituent minerals. Three of the specimens are unsheared and represent stages in a continuous variation from a one-pyroxene (subcalcic ferroaugite) dolerite through partially amphibolized two-pyroxene dolerite to garnet–plagioclase amphibolite. These changes are not accompanied by conspicuous changes in bulk composition. The garnetiferous varieties represent a border facies of the dyke, which is in contact with little-altered pyroxene-bearing gneisses of the Lewisian complex. The fourth specimen is a hornblende schist formed from the dolerite by shearing. This change also involves little change in bulk composition, and affects impartially the three types of dolerite referred to above. Mineral relations are discussed and a hypothesis advanced to account for the garnetiferous border facies of the intrusion, invoking intrusion while the country rock was at relatively high temperature and pressure.

THE petrology of the dyke at Scourie was first described by Teall (1885) who demonstrated that in its least altered state the dyke was composed of an iron-rich quartz-dolerite containing one pyroxene, and that this dolerite was converted to hornblende schist without major change in bulk composition. This conversion was ascribed to dynamic metamorphism at 'ordinary' temperature, a view which was later modified (Teall, 1918) to dynamic metamorphism at 'comparatively low' temperature, i.e. less than that required for fusion. Detailed accounts of the field relations and regional setting of the Scourie dyke and other similar dykes in Sutherland and Wester Ross have been published elsewhere (Clough, Teall, *et al.*, 1907; Sutton and Watson, 1951 *a* and *b*) and need not be recapitulated here.

The clouded feldspars of the Scourie dyke have been commented on by Macgregor (1931) and several aspects of their petrography were examined by Bailey (1951). Four facts not recorded by previous workers are: first, that garnet is abundant in the Scourie dyke at both localities where it is well exposed on the north shore of Scourie Bay (Creag a'

Mhail and Poll Eorna on the 6-in. map, national grid references 29/145458 and 29/150455 respectively); the garnet is restricted to the finer-grained border facies of the dyke, and is absent from any rock that has suffered shearing.

Secondly, undisturbed junctions between dyke and country rock are exposed at Creag a' Mhail (northern contact immediately west of the only convenient access to the dyke hollow), and Poll Eorna (northern contact under shingle near high-water mark a few feet west of a well-exposed bollard of country rock). The dolerite at the contact is a fine-grained rock with a layering of light and dark components parallel to the junction, suggesting flow-banding. This layering persists for a few inches only into the dyke, and is due to segregation of garnet and amphibole. The country rock is a quartz–andesine–augite–hypersthene gneiss that shows no evidence of shearing, even in contact with the layered dolerite. The mineralogy and attitude of the gneiss demonstrate that it belongs to the early Lewisian metamorphic complex with little or no modification due to the late Lewisian metamorphism (Sutton and Watson (1951*a*)).

Thirdly, all of the southern contact and most of the northern contact are affected by shearing, which also affects parts of the centre of the dyke within narrow transgressive zones. Within these zones the dolerite is everywhere converted to a lineated hornblende schist of uniform appearance. At Creag a' Mhail the shearing is seen to be later than the formation of garnet in the border facies. Sutton and Watson (1951*a*) have correlated the shearing with the period of the late Lewisian metamorphism on reliable evidence.

Fourthly, no unmodified chilled contact exists between the Scourie dyke and the gneiss, nor can any normal chilled contact be found between any member of the swarm of similar dykes and the gneiss anywhere in the region between Laxford and Lochinver, nor in Torridon.

Petrology.

One-pyroxene dolerite [84249] (numbers in brackets refer to catalogue numbers in the Harker Collection, Dept. of Mineralogy and Petrology, Cambridge). This material is comparable to that described, analysed, and figured by Teall (1885). The only source of this material appears to be the branch that diverges to the south of the main dyke immediately east of the beach at Poll Eorna. An analysis and mode of this rock are given in table I. The norm yields: Qu 0·1, Or 4·4, Ab 20·3, An 23·8, Di 18·4, Hy 23·5, Il 4·6, Mt 3·8, Ap 0·4, Rest 0·7. Mode and norm agree closely

TABLE I. Chemical and modal analyses of rocks from the Scourie dyke.

	1	2	3	4	5	6
SiO_2	48·58	47·45	48·26	48·23	48·30	49·78
TiO_2	2·43	1·47	2·56	2·96	2·83	2·22
Al_2O_3	13·48	14·83	13·52	13·22	13·34	13·13
Fe_2O_3	2·62	2·47	2·65	4·56	3·93	4·35
FeO	14·07	14·71	13·15	12·62	12·45	11·71
MnO	0·22	n.d.	0·24	0·30	0·21	0·27
MgO	5·11	5·00	5·41	5·20	5·43	5·40
CaO	9·48	8·87	10·11	8·61	9·52	8·92
Na_2O	2·40	2·97	2·35	1·97	2·40	2·39
K_2O	0·75	0·99	0·66	0·96	0·73	1·05
P_2O_5	0·23	n.d	0·12	0·04	0·23	n.d.
H_2O^+	0·55	1·00	0·89	1·53	0·76	1·14
H_2O^-	0·24		0·04	0·12	0·12	
CO_2	n.d.	0·36	n.d.	n.d.	n.d.	0·10
Total	100·16	100·12	99·96	100·32	100·25	100·46
	Modal analyses (vol. %)					
Clinopyroxene	43·6	—	24·7	6·0	—	
Orthopyroxene	—	—	6·4	—	—	
Amphibole	5 7	—	25·6	47·3	70·4	
Plagioclase	40·9	—	34·8	24·2	22·7	
Quartz	3·6	—	2·3	—	—	
Orthoclase						
Garnet	—	—	—	15·0	—	
Biotite	—	—	—	0·6	—	
Iron ore	5·2	—	5·2	6·9	5·7	
Sphene	—	—	—	—	0·6	
Apatite	1·0	—	1·0	tr.	0·6	
	Trace elements (parts per million)					
Cr	100	—	100	40	45	
Ni	80	—	45	30	40	
Co	45	—	45	30	45	
V	320	—	500	450	450	
Li	3	—	6	22	10	
Sr	125	—	100	650	125	
Ba	220	—	160	150	240	
Rb	25	—	30	30	20	
Ga	15	—	20	15	15	
Yt	45	—	25	50	45	
Zr	220	—	140	220	140	
Sc	45	—	55	45	45	

1. 84249 Fresh dolerite, Poll Eorna.
2. Fresh dolerite, Scourie dyke (Teall, 1885).
3. 84248. Altered dolerite, Poll Eorna.
4. 84059. Marginal garnet-plagioclase amphibolite, Creag a' Mhail.
5. 84218. Hornblende schist, Creag a' Mhail.
6. Hornblende schist, Creag a' Mhail (Teall, 1885).

Analyst: 1, 3, 4, and 5, M. J. O'Hara. Trace element determinations by R. Allen, Department of Mineralogy and Petrology, Cambridge.

with the exception that in the mode normative hypersthene appears together with diopside as a single clinopyroxene. The rock is a medium-grained non-porphyritic unfoliated dolerite. In thin section both the

pyroxene and the plagioclase are seen to be clouded by multitudes of small inclusions.

The plagioclase forms tablets that are often partly enclosed by the pyroxene. Zoning is apparent, the range of which has been determined by universal stage measurements (Turner, 1947) as from cores of labradorite (An_{63}) to margins at least as sodic as andesine (An_{36}). Reversals of zoning are rare. Albite and pericline twinning are frequent, Carlsbad and Manebach laws are rarely observed. Many of the tablets are bent. The clouding is due to very fine particles, which are concentrated towards the grain centres, and more particularly into the (010) twin planes. Particles vary in size, and the largest can just be resolved as stumpy rod-like forms of pale green material with high relief, perhaps amphibole.

The pyroxene forms euhedral prismatic crystals. Clouding by what appear to be minute opaque particles is general, but varies in intensity in zones concentric about [001] and is most conspicuous in the outer parts of the grains. In sections parallel to (010) the particles are parallel to the $\{h0l\}$ trace, the angular relationship suggesting that they may be parallel to (001). Twinning on (100) about the normal to (100) is common. The optical properties are: optic axial plane $\|$ (010); $2V_\gamma$ 10° (cores) to 46° (margins) extreme limits of zoning; β 1·711, no detected variation with zoning; γ:[001] 45°; very slight pleochroism in shades of pale brown. An analysis (table II) of the separated zoned crystals shows this to be a sub-calcic ferroaugite, which is probably zoned from pigeonite cores towards ferroaugite margins as indicated by the optical data.

X-ray single-crystal examination by the method of Bown and Gay (1959) showed that the pyroxene is an intergrowth of two clinopyroxenes in approximately equal proportions, the two phases having (001) and [010] in common. It is probable that the two phases are ferropigeonite and ferroaugite. No reflections corresponding with orthopyroxene, amphibole, iron ore, or spinel were found. Similar photographs, differing only in the unit-cell dimensions of the phases, were obtained from almost unclouded and intensely clouded fragments, suggesting that the clouding is due to the exsolution of two phases from the original subcalcic ferroaugite, and that variation in size of the exsolution bodies is the predominant cause of variation in intensity of the clouding.

The iron ore forms skeletal masses and is composed of ilmenite with subordinate magnetite, which is intergrown with the ilmenite as coarse lamellae parallel to one extinction direction of the ilmenite. Alkali feldspar, quartz, and apatite occur interstitially between the plagioclase

TABLE II. Analyses of minerals from the Scourie dyke.

	7	8	9	10	11	12	13	14	15
SiO_2 ...	47·80	48·46	48·33	48·02	44·54	51·51	41·73	36·75	41·25
TiO_2 ...	1·04	0·75	0·83	0·97	1·00	0·92	1·55	1·37	2·43
Al_2O_3 ...	4·65	5·34	4·48	2·49	9·18	1·90	11·85	21·09	13·02
Fe_2O_3 ...	1·75	1·24	2·23	2·24	5·07	2·87	5·98	4·10	4·14
FeO ...	22·42	18·15	18·51	30·78	17·82	9·26	14·00	25·12	15·94
MnO ...	0·42	0·12	0·26	0·62	0·28	0·13	0·11	1·52	0·26
MgO ...	9·66	10·78	10·60	12·33	7·64	11·41	8·13	2·39	7·22
CaO ...	11·31	14·03	14·07	2·02	11·03	21·91	11·63	8·12	11·30
Na_2O ...	0·51	0·62	0·63	0·22	1·20	0·58	1·39	n.d.	1·64
K_2O ...	0·28	0·28	0·25	0·18	1·26	0·05	1·09	n.d.	0·92
P_2O_5 ...	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0·05
H_2O^+ ...	n.d.	0·26	n.d.	n.d.	1·35	n.d.	2·03	n.d.	0·85
H_2O^- ...	n.d.		n.d.	n.d.	nil	n.d.	0·09	n.d.	0·07
Total ...	99·84	100·03	100·19	99·87	100·37	100·54	99·58	100·46	100·09
Si ...	7·47	7·44	7·46	7·61	6·80	7·74	6·32	5·81	6·28
$Al^{[4]}$ ...	0·53	0·56	0·54	0·39	1·20	0·26	1·68	3·92	1·72
$Al^{[6]}$ ...	0·35	0·41	0·27	0·07	0·45	0·08	0·43		0·61
Fe^{3+} ...	0·20	0·18	0·26	0·27	0·59	0·32	0·68	0·49	0·47
Fe^{2+} ...	2·92	2·32	2·38	4·08	2·27	1·16	1·77	3·31	1·93
Ti ...	0·12	0·08	0·10	0·12	0·12	0·10	0·18	0·16	0·28
Mn ...	0·06	0·02	0·03	0·08	0·04	0·02	0·01	0·20	0·03
Mg ...	2·24	2·46	2·43	2·91	1·73	2·55	1·83	0·56	1·63
Ca ...	1·90	2·31	2·32	0·34	1·80	3·53	1·88	1·38	1·84
Na ...	0·16	0·18	0·19	0·07	0·36	0·17	0·41	—	0·18
K ...	0·06	0·06	0·05	0·04	0·25	0·01	0·21	—	0·06
O ...	24·0	24·0	24·0	24·0	24·0	24·0	24·0	24·0	24·0
H ...	—	—	—	—	1·37	—	2·05	—	1·86
Trace elements (parts per million)									
Cr ...	120	180	—	40	75	220	80	5	45
Ni ...	70	35	—	45	45	32	22	*	40
Co ...	60	60	—	60	45	55	30	*	45
V ...	550	600	—	220	500	600	1000	220	1000
Li ...	12	22	—	3	16	20	22	2	10
Sr ...	10	*	—	*	5	30	22	*	25
Ba ...	65	45	—	200	150	5	150	45	200
Rb ...	*	*	—	*	30	*	*	*	20
Ga ...	8	10	—	15	15	10	20	15	15
Yt ...	45	45	—	*	50	140	45	*	25
Zr ...	120	100	—	100	120	100	45	120	120
Sc ...	100	120	—	*	70	*	100	120	45

* Below limit of sensitivity.

	Opt. data, page
7. Subcalcic ferroaugite from fresh dolerite. 84249.	851
8 and 9. Ferroaugites from altered dolerite. 84248.	854
10. Ferrohypersthene from altered dolerite. 84248.	
11. Amphibole from altered dolerite. 84248.	853
12. Augite from garnet–plagioclase amphibolite. 84059.	855
13. Amphibole from garnet–plagioclase amphibolite. 84059.	
14. Garnet from garnet–plagioclase amphibolite. 84059	
15. Amphibole from hornblende schist. 84218.	856

Analyst: M. J. O'Hara. Trace element determinations by R. Allen.

tablets, and represent a late crystallized portion of the magma. The alkali feldspar has $2V_{\alpha}$ 49°, α 1·522, γ 1·529, consistent with a high-temperature-state orthoclase of composition about $Or_{60}Ab_{40}$ (Tuttle, 1952). Although optically homogeneous, an X-ray single-crystal study (method of Mackenzie and Smith, 1955) shows this to be an intergrowth of subsidiary amounts of albite-twinned sodic feldspar and dominant orthoclase. The orientation of the two components is that expected in an exsolution intergrowth, and the material is presumably cryptoperthitic.

Amphibole is present in very small amounts, as rims about iron ore and pyroxene. It is never euhedral and appears to be secondary. Minor scapolite and sericite are present as alteration products of plagioclase. A few samples from this locality show less amphibole and less conspicuous clouding of pyroxene and plagioclase than the specimen described.

Two-pyroxene dolerite [84248]. Similar material may be collected from the centre of the main dyke at either locality, the specimen described here coming from Poll Eorna. An analysis and mode are given in table I. Although chemically very similar to the one-pyroxene dolerite there are five principal distinctions: a higher combined water content and higher Fe_2O_3/FeO ratio are noted; this is not connected with a higher content of modal iron ore. Clouding of plagioclase is less conspicuous due to the presence of fewer rod-like inclusions, which are, however, larger than those described from the one pyroxene dolerite, but otherwise similar.

Thirdly, amphibole is abundant, mantling all pyroxene and iron ore grains; the optic properties are α yellow-green, β 1·691 deep olive, γ 1·699 deep blue-green; an analysis is given in table II. The amphibole occurs as one or more individuals mantling a single pyroxene grain, frequently with the [001] axes of the two minerals parallel. A little iron ore is enclosed within the amphibole, but quartz is not observed in association with these mantles.

Fourthly, the pyroxene grains have the same external form as in the one-pyroxene dolerite, but two pyroxenes are now present. The majority of the material is a pale green clinopyroxene with fibrous or vermicular appearance due to the abundance of lamellae parallel to (100). Twinning on (100) is present but the clouding noted in the pyroxene of the one-pyroxene dolerite is absent. The optical properties of the host are β 1·706, $2V_{\gamma}$ 50°, indicating a composition $Ca_{39}Mg_{32}Fe_{29}$ (Hess, 1949; Tröger, 1956). X-ray single-crystal studies show clinopyroxene with abundant orthopyroxene intergrown with (100) and [001] common to the two structures, the orientation found in intergrowths attributed to exsolution (Bown and Gay, 1959). Some amphibole is present also.

Analyses of two portions of the clinopyroxene differing in density and magnetic properties are given in table II. These are bulk compositions representing the host ferroaugite plus the orthopyroxene lamellae.

Also present are patches of orthopyroxene varying in form from coarse blebs enclosed in the augite and elongated parallel to the trace of (001) of the augite through ragged areas marginal to large augite grains to apparently independent crystals. The optical properties are α 1·726 pale red, β yellow-green, γ 1·737 apple green, indicating a composition $Mg_{40}Fe_{60}$ (Hess, 1952). Inclusions are apparent in thin section and X-ray single-crystal study shows the presence of substantial amounts of clinopyroxene oriented with (100) and [001] in common with the orthopyroxene host, together with additional reflections due to clinopyroxene in no special orientation. The latter are similar to those found in inverted pigeonites (Bown, personal communication). An analysis of the orthopyroxene plus the intergrown clinopyroxene is given in table II. Both pyroxenes contain small translucent plates of a brown material that has not been identified.

Fifthly, the alkali feldspar is microperthitic, containing abundant oriented spindles of albite. The host is orthoclase, with $2V_{\alpha}$ 52°, α 1·520, γ 1·526 indicating a composition $Or_{75}Ab_{25}$ (Tuttle, 1952). An X-ray single-crystal study of a microperthite from a neighbouring specimen gave a similar result to that for the alkali feldspar of the one-pyroxene dolerite.

The ore mineral in this rock is ilmenite with subsidiary magnetite. Lamellar intergrowth is not found, but the distribution of magnetite in some of the granular aggregates suggests recrystallization of an original lamellar intergrowth.

Garnet plagioclase amphibolite [84059]. This specimen, collected from a point 3 in. within the dolerite at the northern contact, Creag a' Mhail, is a fine-grained massive unfoliated rock with a patchy appearance in hand specimen due to the irregular distribution of pale garnet. An analysis and mode are given in table I, whence it is apparent that the mineralogical differences are not associated with changes of chemical composition other than an increase in combined water and a rise in Fe_2O_3/FeO relative to both types of dolerite described above.

In thin section the texture is seen to have been derived by alteration of a holocrystalline fine-grained dolerite. The form of plagioclase tablets is preserved, although these are much altered, and encroached on by the aggregates of fine-grained amphibole and quartz that pseudomorph the pyroxene. A few relics of fibrous clinopyroxene similar to that described

from the two-pyroxene dolerite are present, but these are commonly recrystallized to a fine granular aggregate of clear pale green clinopyroxene that has β 1·698, $2V_\gamma$ 58° indicating a composition $Ca_{46}Mg_{34}Fe_{20}$ (Hess, 1949) in close agreement with that found by analysis (table II). The clinopyroxene is present in accessory amounts only, and it is associated intimately with quartz, but not directly with amphibole or garnet.

Garnet occurs as large (2 to 3 cm.) clusters of small (0·02 cm.) euhedral grains accompanied by abundant iron ore and quartz and lesser amounts of amphibole. The refractive index is n 1·803, and the cell-size a 11·603 Å. An analysis is given in table II confirming that this is an almandine-rich variety.

The amphibole has α 1·669 yellow-green, β 1·680 olive, γ 1·690 blue-green, γ:[001] 18°. It occurs with quartz, iron ores, and plagioclase throughout the body of the rock. An analysis is given in table II.

A few small anhedral grains of apatite are present, but alkali feldspar is absent. The recrystallized granular plagioclase is untwinned and difficult to distinguish from quartz. A refractive index measurement suggests a composition close to $Ab_{65}An_{35}$. The ore minerals are present only as small grains. An X-ray powder photograph of the separated ore minerals shows the presence of magnetite, ilmenite, and subordinate hematite. No intergrowths are apparent. A little red-brown biotite is present as small flakes.

The grain size of the minerals in this rock is of the order of 0·02 cm., in contrast with the 0·15 cm. grain size of the fine-grained dolerite from which it appears to have been derived. This dolerite was finer grained than the pyroxene-bearing dolerites from the centre of the dyke which have a grain size of 0·2–0·4 cm.

Samples [84039] and [84040] from Poll Eorna, which show the contact with the gneiss, differ from the specimen just described in: scarcity of clinopyroxene; absence of a texture reminiscent of a holocrystalline dolerite; presence of irregular tablets of plagioclase, most of which are oriented parallel to the contact surface; and presence of layering parallel to the contact due to segregation of garnet and amphibole. The amphibole shows no preferred orientation.

The contact itself is sharp and apparently undisturbed by shearing. The gneiss is a coarse-grained granular acid rock whose features have been described in the section on field observations.

Hornblende schist [84218]. This is a foliated and lineated rock, dark green in hand specimen, which was collected from a shear zone in the

dolerite at Creag a' Mhail and near the centre of the dyke. An analysis and mode are given in table I. In thin section the rock is seen to be completely reconstructed to a medium-grained hornblende–plagioclase–quartz schist that contains prominent schlieren of iron ore. Accessory apatite is present. None of the minerals exhibit good crystal form, most occurring as lozenge-shaped grains elongated parallel to the plane of foliation.

The amphibole is a pleochroic iron-rich variety (analysis, table II) with the scheme: α 1·665 yellow-green, β 1·680 deep olive-green, γ 1·693 deep blue-green, γ:[001] about 28°. Strong absorption prevented satisfactory determination of 2V. There is pronounced preferred orientation of the amphibole with (100)|| the schistosity and [001] axes in parallel alignment. A parting parallel to (100) is well developed in addition to the good prismatic cleavage. Most grains are elongated parallel to [001].

The plagioclase is andesine, composition $Ab_{64}An_{36}$ as determined by universal stage measurements on the rare twinned grains (method of Turner, 1947) and by refractive index measurements. Zoning and clouding are absent. Owing to the relatively large amount of untwinned feldspar present this mineral has not been distinguished from quartz in the mode. Potash feldspar is absent. The iron ore is largely ilmenite with occasional grains of magnetite. The ilmenite is markedly anisotropic in reflected light and has abundant twin lamellae but no inclusions of magnetite. The magnetite contains three sets of narrow lamellae of ilmenite similar to those ascribed to exsolution when observed in gabbros and dolerites.

Scarce sphene occurs associated with the ilmenite. Garnet is absent from this and all similar hornblende schists that have been examined. Cataclastic effects are similarly lacking, and even the bending of grains noted in the unsheared dolerite is absent from this rock.

Contact facies of Scourie-type dykes.

Further information on the origin of the garnetiferous border facies of the Scourie dyke is provided by an examination of the contacts of other dykes in the same region. Two narrow veins of dolerite 6 to 9 in. wide cut basic pyroxene-granulites on the headland west of Geodh Eanruig (map reference 29/141441). A cave has been eroded through the headland along the line of these dykes, which appear to be composed of a very fine-grained dolerite. Abundant small phenocrysts of plagioclase are visible in hand specimens. In thin section the rock is seen to be composed

of numerous tabular phenocrysts of plagioclase and rarer phenocrysts of clinopyroxene, both minerals being extensively clouded by opaque inclusions. Individual phenocrysts are up to 0·2 cm. in length. The groundmass is a very fine-grained unfoliated granular aggregate of amphibole, iron ores, quartz, untwinned plagioclase, and garnet. The small garnets are gathered into clusters of several hundred grains which give rise to pale globular patches in the hand specimen. Near the contact the phenocrysts of plagioclase are aligned parallel to the junction. The contact is sharp and undisturbed, and the country rock is a coarse-grained granular augite–hypersthene–andesine–antiperthite gneiss, which has all the characteristics of the least altered 'charnockitic' gneisses of the early Lewisian metamorphic complex.

This rock is interpreted as a chilled porphyritic dolerite whose groundmass may originally have been glassy, or a very fine grained material in which alignment of phenocrysts had taken place by flow. The state of the groundmass can best be explained on the assumption that the dyke and its country rock were held under conditions of temperature, pressure, and water-vapour pressure such that the chilled dolerite recrystallized to a mineral assemblage characteristic of the almandine–amphibolite mineral facies, while the granulite facies mineralogy of the country rocks persisted metastably. This recrystallization must have preceded the period of the late Lewisian metamorphism because in this narrow dyke, as in the Scourie dyke, the garnetiferous dolerite is converted to hornblende schist within narrow shear belts.

The results of examination of the contacts of dykes similar to the Scourie dyke throughout the Scourie–Lochinver region are presented in table III and fig. 1. The extent to which the textural evidence of original chilling has been obscured in the dyke contacts increases with the width of the dyke concerned, but is independent of the position of the dyke relative to the areas affected by the late regional metamorphism. Even in the broadest dykes the presence in contact specimens of aligned but corroded plagioclase phenocrysts in a matrix of unfoliated amphibolite or garnet amphibolite still affords evidence of original chilling. The occurrence of garnet in the border facies of the dykes is most frequent near Scourie, and decreases in a south-easterly direction, i.e. roughly perpendicular to the strike of the foliation in the early complex gneisses. Irrespective of position, however, the recrystallized assemblages belong to the amphibolite mineral facies.

Superimposed on the recrystallization of the borders of the dykes are three later effects: shearing during the late regional metamorphism;

TABLE III. Contact phenomena of Scourie dykes.

Key to observations

1g chilled margin identified, with garnet and hornblende.
1a chilled margin identified, with hornblende only.
2 margin sheared to hornblende schist } Laxfordian events.
3 dyke regionally metamorphosed }
4 pseudotachylyte formed in dyke near margin. Pre-Torridonian.
E epidote present in margin of dyke.
5 epidote veining present.

Map reference	Locality	Observations
1. 29/241476	½ mile NE. of Laxford Bridge	3
2. 29/230445	1 mile SSE of Badnabay	3
3. 29/207461	½ mile SE. of Loch na Claise Fearna	3+4
4. 29/148464	Coast ⅝ mile N. of Creag a' Mhail	2
5. 29/147463	Coast ½ mile N. of Creag a' Mhail	1a+2
6. 29/166452	Near summit of Creag a' Bhadaidh Dharaich	2+4
7. 29/145458	Creag a' Mhail, Scourie Bay	1g+2
8. 29/150455	Poll Eorna, Scourie Bay	1g+2
9. 29/149448	Near graveyard, Scourie Bay	1g+2+4
10. 29/161435	¼ mile N. of Loch an Daimh Mhor	2
11. 29/141441	Promontory W. of Geodh Eanruig. 9i n. wide dyke	1g+2
12. 29/141441	Promontory W. of Geodh Eanruig. 0 to 10 ft. wide dyke	1g+2
13. 29/146438	½ mile WSW. of school, Scourie	2+4
14. 29/146417	½ mile NW. of Farhead point, Badcall	1a+2
15. 29/197392	W. shore of Loch Crocach	2
16. 29/195388	S. end of Loch Crocach	1a+4
17. 29/194385	⅓ mile S. by W. of Loch Crocach	2
18. 29/194384	,, ,, ,, ,,	1g+2
19. 29/196380	⅓ mile S. of Loch Crocach	1g+2
20. 29/245342	½ mile W. of mouth of Maldie Burn	2
21. 29/261358	Near S. end of Loch na Leathaid Bhuain	1g
22. 29/272367	Allt Beallach a' Phollaidh	3+4
23. 29/255338	½ mile E. of Maldie Burn	1a+2+4+E
24. 29/263337	¾ mile E. of Maldie Burn	1a+2+E
25. 29/268338	1 mile E. of Maldie Burn	1a+E
26. 29/274339	½ mile W. of Glendhu house	2+E
27. 29/074211	Badanaban	2+4+E
28. 29/081207	,,	1a
29 29/075212	,,	1a+2+E
30. 29/096233	NE. of Lochinver bridge	1a+2+E
31. 29/113239	Lochinver-Inchnadamph road	1a+E
32. 29/123240	N. of large ultrabasic dyke at Brackloch	2
33. 29/198262	1 mile ESE. of Tumore, Loch Assynt	1a+2+E
34. 29/285234	½ mile SW. of Loch nan Ouaran	1a+2+5
35. 29/285235	,, ,, ,, ,,	1a+2+5
36. 29/285223	¼ mile W. of Lagan Mhuirich	1a+2+5
37. 29/285226	,, ,, ,, ,,	1a+2+5
38. 29/274303	Glencoul Lodge	3+5
39. 29/303224	½ mile W. of Garbh Allt	1a+5
40. 29/126158	Near W. end of Loch Veyatie, Assynt	1a+2
41. 29/133159	,, ,, ,, ,, ,, ,, ,,	1a+2+4
42. 29/138159	,, ,, ,, ,, ,, ,,	1a+2
43. 29/142156	,, ,, ,, ,, ,, ,, ,,	1a+2
44. 29/143157	,, ,, ,, ,, ,, ,, ,,	1a+2
45. 29/145157	,, ,, ,, ,, ,, ,, ,,	1a+2

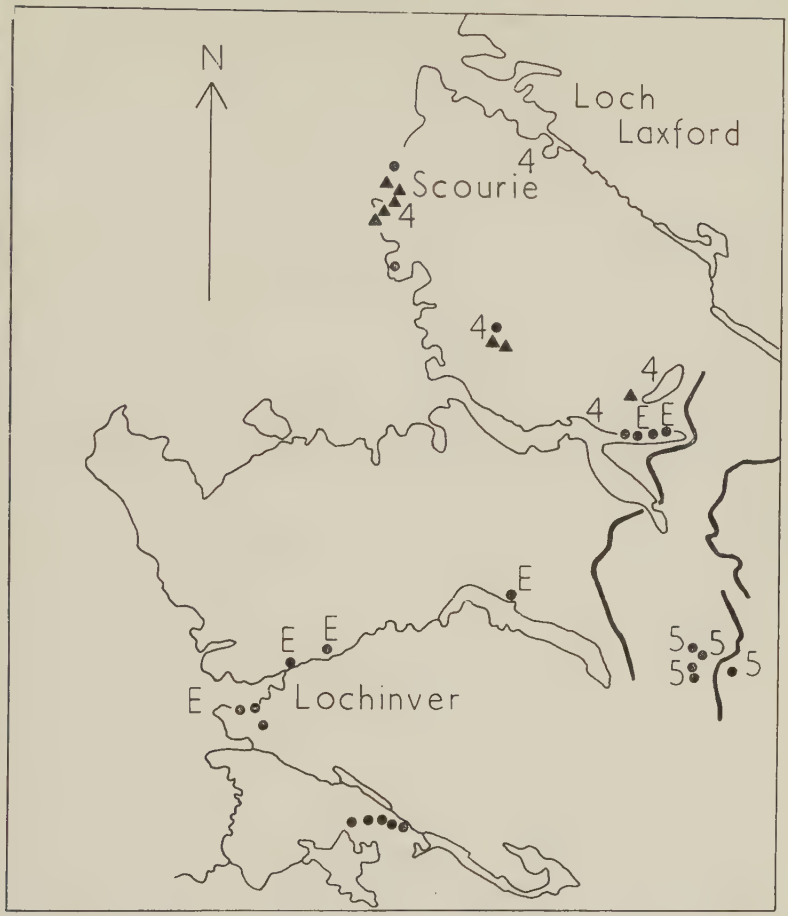


FIG. 1. Sketch-map of the Scourie–Assynt district showing the location of recognizable chilled contacts to Scourie dykes. Those containing garnet as well as amphibole are shown by triangles, those with amphibole only by circles. The situations of pseudotachylyte (4), epidote (E), and epidote vein (5) occurrences are also shown relative to the outcrops of the principal post-Cambrian thrust planes which are shown by heavy lines.

crushing with the formation of pseudotachylyte; and crushing and low-grade metamorphism, with the formation of chlorite and epidote, which is associated with proximity to the post-Cambrian thrust planes.

The varying extent of recrystallization of the chilled margin can only be accounted for if it is supposed that the factor producing the recrystallization of the contacts is dependent upon the dyke width. The heat

capacity of the magma is the most obvious factor that would vary in this manner. Accordingly it is proposed that the recrystallization took place immediately after intrusion and because the dykes were intruded into the gneiss at a time when the rocks were still deeply buried and relatively hot. (A temperature of 300–500° C. would be consistent both with chilling of the magma and with the generation of amphibolite facies assemblages.) Recrystallization is more advanced in the border zones both because these chilled assemblages were further from equilibrium under the imposed conditions, and because the border zones were subject to an auto-thermal metamorphism by the heat-flow from the cooling interior.

The petrology of the Scourie dyke is consistent with the crystallization of a one-pyroxene amphibole-free dolerite as the primary product, this material being modified towards a mineral assemblage stable under the external pressure and temperature conditions by an amount dependent upon the distribution of volatiles and the subsequent heat flow. The extent of modification is thus loosely correlated with distance from the centre of the dyke.

Mineralogical changes during metamorphism of the Scourie dyke.

The pyroxenes of the Scourie dyke samples are considered in relation to other igneous and metamorphic pyroxenes in fig. 2. The modification of the primary high-temperature phase by exsolution is complicated by the accompanying incorporation of some of the pyroxene into the amphibole. The amphibole has a higher Fe:Mg ratio (fig. 3) than the associated pyroxene and preferentially takes up the calcium-poor component of the pyroxene. The sequence of events leading towards the attainment of a stable amphibolite facies mineralogy is: first, crystallization of a zoned subcalcic ferroaugite.

Secondly, exsolution of ferropigeonite and consequent increase in the calcium content of the subcalcic ferroaugite. This is accompanied by the appearance of an iron-rich amphibole by reaction between pyroxene and plagioclase, with resulting decrease in the Fe:Mg ratio of the bulk pyroxene assemblage.

Thirdly, inversion of the ferropigeonite to ferrohypersthene, accompanied by further exsolution of ferroaugite from the ferrohypersthene and of ferrohypersthene from the ferroaugite, which has now attained a relatively high calcium content. An accompanying increase in the volume of amphibole leads to a further decrease in the Fe:Mg ratio in the pyroxenes, further reduction in the total volume of pyroxene, and

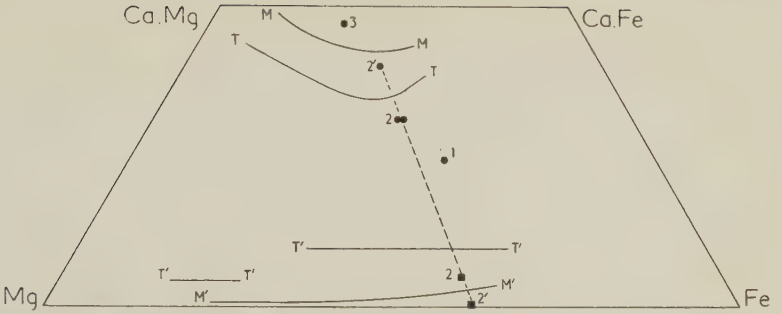


FIG. 2. Part of the Ca–Mg–(Fe+Mn) diagram. T, T′ are curves representing the compositions of coexisting calcium-rich and calcium-poor pyroxenes from tholeiitic gabbros (Brown, 1957; Muir, 1954). M, M′ are similar curves for coexisting pyroxenes from the Madras charnockite series (Howie, 1955). The numbered points represent data from the Scourie dyke: 1, bulk composition of the primary zoned pyroxene of the fresh dolerite; 2, bulk compositions of two portions of clinopyroxene and of the coexisting orthopyroxene from the altered dolerite; each phase contains lamellae of the other and 2′ represents the composition of the host phases (derived from the optical properties). 3 is the composition of the accessory augite from the garnet-plagioclase amphibolite at the margin of the dyke.

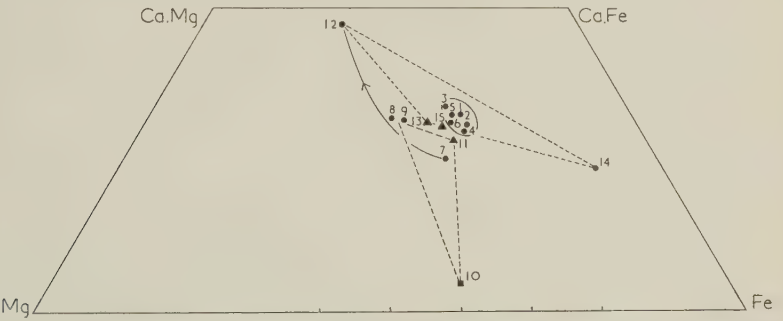


FIG. 3. Compositions of analysed Scourie dyke samples compared in the diagram Ca–Mg–(Fe+Mn). 1 to 6 are the rock compositions numbered to correspond with the analyses in table I. A field for the composition of samples from this dyke is suggested. Other points are numbered to correspond with the mineral analyses of table II: 7, primary pyroxene of fresh dolerite; 8, 9, clinopyroxenes, 10, orthopyroxene, and 11, amphibole of the altered dolerite; 12, clinopyroxene, 13, amphibole, and 14, garnet of the marginal garnet-plagioclase amphibolite. 15, amphibole from the hornblende schist. Broken lines link coexisting minerals (not necessarily in equilibrium), and a curved line with arrow indicates the inferred trend of the bulk pyroxene composition during amphibolization.

selective reduction in the volume of ferrohypersthene relative to ferroaugite.

Finally, further increase in the amount of amphibole leads to the modification of the residual pyroxene, now exclusively clinopyroxene,

to a calcium-rich augite of relatively low Fe:Mg ratio. Garnet appears in abundance.

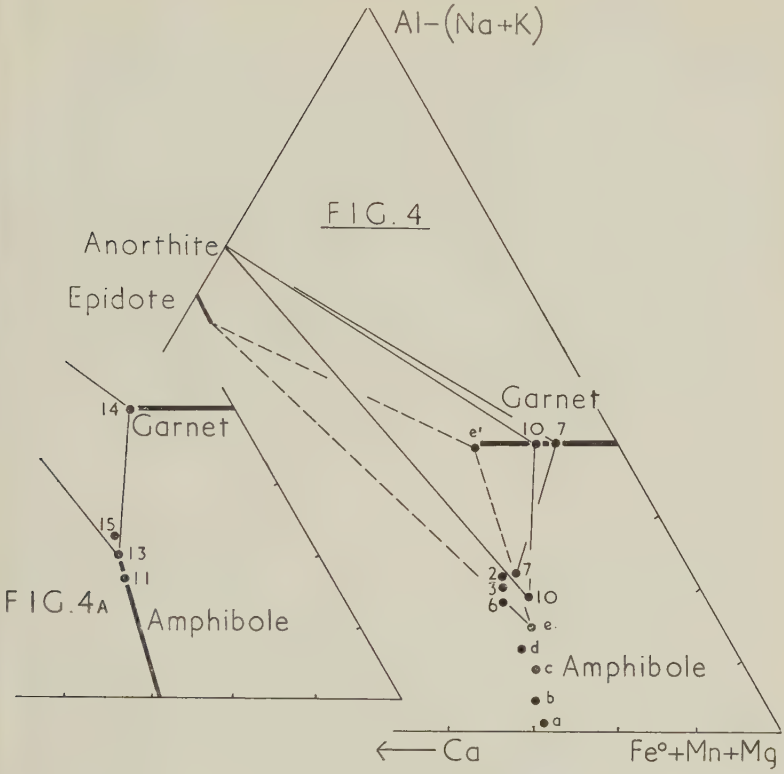
A parallel series of changes is exhibited by the alkali feldspar, which undergoes exsolution followed by incorporation of the potassium in the amphibole.

The submicroscopic exsolution and clouding in the pyroxene and feldspars of the one-pyroxene dolerite need not belong to the immediate post-intrusion period, and may have occurred during the period of the late regional metamorphism. However, clouding of the plagioclase appears to vary with the other aspects of the mineralogy and it is at least possible that the clouding is the result of the rocks being held at a comparatively high temperature for a period immediately following their intrusion as has been suggested in some other cases (Poldervaart and Gilkey, 1953). This does not invalidate Macgregor's (1931) conclusion that the clouding is due to subsequent thermal metamorphism.

Amphibole- and garnet-bearing parageneses are considered in fig. 4, where a case is made for considering the aluminium content (as represented by Al–[Na + K]) of amphibole as a function of metamorphic grade, and for considering the appearance of garnet within the amphibolite facies as an indication that the amphibole possesses the maximum aluminium substitution under the particular conditions. The role of garnet is to contain any further aluminium over that which can be accommodated by feldspar+amphibole. On this basis the appearance of garnet within a regional sequence of amphibolite facies metadolerites is an indication either of decreasing metamorphic grade or of a regional variation in bulk composition of the rocks concerned (increase of normative hypersthene relative to diopside). Reference to fig. 1 leads to the conclusion that either the physical conditions within the early complex varied from north-west to south-east at the time of intrusion of the dykes similar to the Scourie dyke, with the higher grade conditions to the south-east, or there is a regional variation in bulk composition in the same direction.

In fig. 4A the analysed amphiboles and garnet from the Scourie dyke are compared. The low aluminium content of the amphibole from the two-pyroxene dolerite is consistent with the absence of garnet from this rock. The aluminium content of the amphibole from the hornblende schist formed during the late regional metamorphism is appreciably greater than that of the unsheared border facies of the dyke.

This might be taken as evidence of conditions of higher temperature or pressure or both in the shear zones during the later metamorphism



FIGS. 4, 4A: FIG. 4. Ternary plot (Ca; Al (Na + K); total iron + Mn + Mg) of amphibole and garnet analyses from metadolerites. The points *a*, *b*, *c* represent amphiboles coexisting with aluminous chlorite in three greenschist facies rocks (Tilley, 1938). *d*, *e* represent amphiboles from epidote-amphibolite facies epidiorites at Achahoish (Wiseman, 1934). The amphibole *e* is accompanied by epidote and a garnet *e'*. 2, 3, 6 represent amphiboles from garnet-free plagioclase amphibolites of the Adirondack region (upper amphibolite facies) and 7, 10 are coexisting amphiboles and garnets from garnet–plagioclase amphibolites of the same region; numbers refer to analysed assemblages in Buddington, 1952. The lines link amphibole, garnet, and the appropriate calcium aluminosilicate. FIG. 4A. Part of a ternary plot, as in fig. 4, of amphiboles and garnets from the Scourie dyke. 11, amphibole of altered dolerite; 13, amphibole, 14, garnet from marginal garnet–plagioclase amphibolite, and 15, amphibole from hornblende-schist. Tie lines for feldspar–amphibole garnet of the marginal garnet–plagioclase amphibolite are shown. Numbers refer to analyses in table II.

than were prevalent at the intrusion of the dykes, and might indicate development of the pressure and temperature conditions of the amphibolite facies throughout the Lewisian complex at the time of the later

metamorphism, with reconstruction to amphibolite-facies mineral assemblages taking place only in zones affected by local or regional shearing and metasomatism.

Within the shear zones in the early complex both gneisses and dykes were converted directly to an amphibolite facies mineralogy and transition from the early complex into the late metamorphic complex also involves the replacement of the existing mineral assemblages by those of the amphibolite facies (Sutton and Watson, 1951 *a*, *b*) without intervening zones of lower grade rocks.

Noë Nygaard (1948) in a closely comparable instance from Greenland has recorded the development of a zone of low-grade rocks between an early and a late metamorphic complex. The dykes that cut and chill against the early complex in Greenland and are metamorphosed in the late complex, have normal contact features (Sørensen, personal communication).

Acknowledgements. I wish to thank Prof. C. E. Tilley for suggesting the subject for study, Dr. S. R. Nockolds for permission to reproduce trace element determinations, Prof. Tilley, Dr. Nockolds, Dr. S. O. Agrell, and Dr. J. Phemister for their advice and criticism, and Prof. F. H. Stewart for reading the manuscript. I also wish to thank Dr. M. G. Bown for assistance in the interpretation of X-ray photographs. My thanks are due to Miss J. Tarrant and Miss A. Gray for their assistance in the preparation of this paper. The study was made in the Department of Mineralogy and Petrology, Cambridge, and was made possible by the award of a maintenance grant by the Department of Scientific and Industrial Research.

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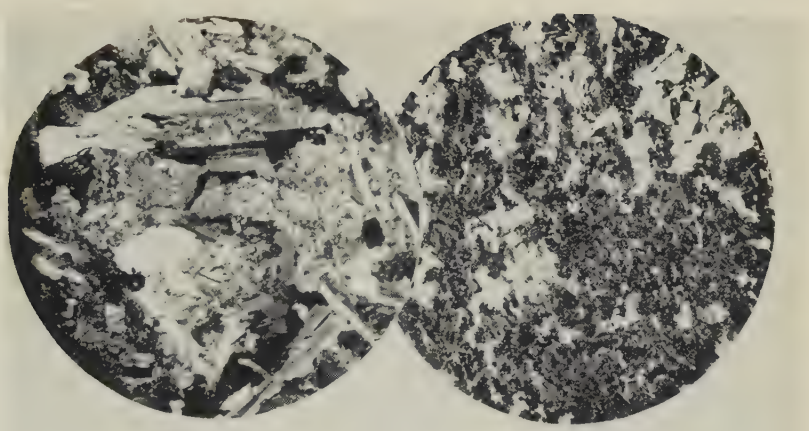

Fig A Fig B

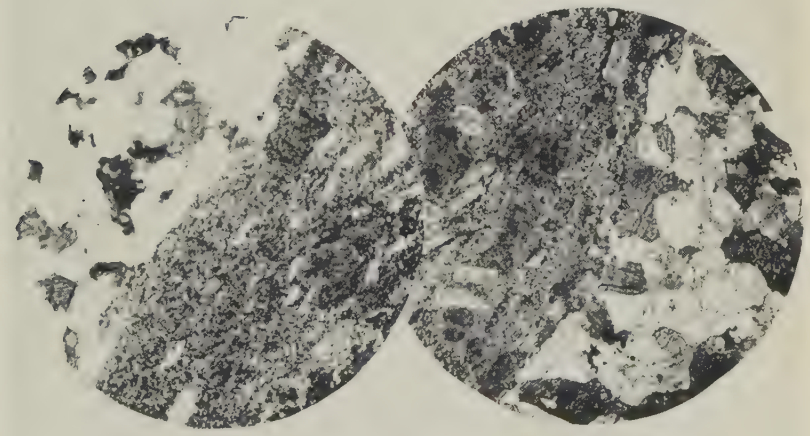

Fig C Fig D

M. J. O'Hara: The Scourie dyke

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Explanation of Plate XVIII.

Fig. A. Altered dolerite (84084) from the Scourie dyke at Poll Eorna. At the centre of the field is a composite grain, partly mantled by iron ore, which consists of ferrohypersthene (grey), forming an independent-looking grain at the left but intergrown as blebs with ferroaugite (pale) at the right. The intergrowth is parallel to (100) of both pyroxenes. Crossed nicols, ×8.

Fig. B. Medium-grained chilled dolerite (84219) 4 in. from the northern contact of the Scourie dyke at Poll Eorna. Amphibole and areas rich in small garnets have developed between plagioclase and pyroxene. One nicol, ×8.

Fig. C. Northern contact of the Scourie dyke and acid gneiss from Poll Eorna. Specimen (84040). The coarse-grained gneiss is unsheared but shows partial amphibolization of pyroxenes. The dolerite contains modified tabular phenocrysts of plagioclase oriented parallel to the contact. A garnet-rich area in the plagioclase amphibolite groundmass is apparent at top right. One nicol, ×8.

Fig. D. Contact of 9-in. vein of chilled dolerite and basic pyroxene gneiss from Geodh' Eanruig, Scourie More. Specimen (84055). Tabular phenocrysts of plagioclase are oriented parallel to the contact. Scarcer phenocrysts of clinopyroxene appear dark due to inclusions of iron ore. The groundmass is composed of amphibole, plagioclase, and iron ore with conspicuous clusters of small granules of garnet. One nicol, ×8.

Metamorphism and metasomatism of rocks of the Moine Series by a dolerite plug in Glenmore, Ardnamurchan.

(With Plate XIX.)

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[Read 23 March 1961.]

Summary. A small dolerite plug in Glenmore, Ardnamurchan, has metamorphosed and partially vitrified the adjacent Moine arkoses up to a distance of 6 ft. from the contact. Chemical analyses show that the glasses produced by the metamorphism have compositions close to the ternary minimum in the system $NaAlSi_3O_8$–$KAlSi_3O_8$–SiO_2, indicating that they were formed by fractional fusion of a lowest-melting composition from the arkose. Alkali metasomatism in the metamorphosed arkoses resulted in an increase of Na_2O and a decrease of K_2O as the contact with the dolerite is approached, while an associated pelitic horizon shows an increase of K_2O content. A review of other examples of thermal metamorphism of quartzo-feldspathic rocks by basic intrusions shows that movements of alkalis are very common in this type of environment, and the possible relations of the metasomatism and the fusion are discussed.

THE partial fusion of quartzo-feldspathic and argillaceous rocks is a common feature of high-temperature thermal metamorphism such as occurs in xenoliths and at the contacts of intrusions of hot basic magma. Many examples are recorded of the partial melting of feldspathic quartzites, pelitic rocks, and acid igneous rocks, and a full bibliography of these is given in a recent paper by Wyllie (1961), while Ackermann and Walker (1960) provide another recent description of the same phenomenon. A bibliography of instances of fusion of siliceous sediments by combustion of coal is given by Venkatesh (1952), and an analogous example is the melting of calcareous sediments by combustion of hydrocarbons described by McLintock (1932). Finally, the fusion of siliceous schists and sandstones in Scottish vitrified forts (Childe and Thorneycroft, 1937), and in rocks involved in Bronze Age cremations (Davidson, 1936) may be mentioned as examples of the artificial melting of rocks, while the use of molten slags in industrial processes is too well known to require detailed reference.

In the area of Ardnamurchan to the east of the Tertiary igneous complex that forms the western end of the peninsula, there are at least

fourteen plugs or dykes of basic rock that cause extensive contact alteration of the Moine Schists, into which they are intruded. They are similar to the dykes and bosses of the Moidart area referred to in the Summary of Progress of the Geological Survey for 1930 (Richey *et al.*, 1931, p. 64). Some half-dozen of the intrusions in Ardnamurchan are recorded on the Geological Survey 6-in. to the mile field sheets (the permission of the Director of the Geological Survey for the author to consult unpublished maps is gratefully acknowledged), and the remainder were discovered by the author in the course of an investigation of the Moine Schists of the Ardnamurchan area. The plugs intrude both the psammitic and the pelitic groups of the Moine Schists, and show considerable variation in outcrop from stock-like forms up to $\frac{1}{4}$ mile in diameter to dyke-like intrusions about $\frac{1}{3}$ mile in length and 20 yds. in width. The majority of the metamorphosed Moine rocks show the development of either felsitic or granophyric texture with abundant sericitic alteration (pl. XIX, fig. 1), and the igneous intrusions that caused the metamorphism are large in outcrop area and are also extensively altered. The metamorphic aureole of a dolerite plug in Glenmore, which is the subject of the present investigation, is the only example in Ardnamurchan in which glass has been found in the arkoses metamorphosed at the contact of an intrusion (pl. XIX, fig. 2). The Glenmore plug is also outstanding in being the smallest and least altered of all the intrusions that cut the Moine Schists in Ardnamurchan east of the Tertiary igneous complex.

Field description.

The Glenmore dolerite plug is intrusive into Moine arkoses $\frac{3}{4}$ mile NNE. of the road bridge over the river at Glenmore, Ardnamurchan. It forms a slight prominence on the south-west side of the ridge north-east of the track from Glenborrodale to Lochan na Carraige at a point due west of Lochan nan Caorach. The National Grid reference of the locality on the 1-in. Ordnance Survey sheet number 45 (Seventh Series) is 591638.

The plug forms an irregular hexagonal outcrop, with the south-west and north-east sides parallel with the strike of the Moine arkoses; the maximum length of the plug is 145 ft. (NW. SE.) and the width is 85 ft. The south-western contact of the dolerite, which forms a steep crag-face overlooking the track, dips at 75° to the south-west, thus coinciding with the dip of the arkoses in the area. The contact of the dolerite and the arkose is thoroughly welded, and a thin skin of partly fused arkose adheres to the south-western side of the plug. On the south side of the plug, which is also well exposed, the contact is irregularly

corrugated, suggesting that some bands of the arkose offered less resistance to emplacement of the dolerite than others. A band of pelite 4 to 6 in. thick is interbedded with the arkoses, and the dolerite forms a dyke-like apophysis of the same width extending along the plane of the pelitic band for a distance of about 1 ft. from the contact of the dolerite with the adjacent arkoses. The banding of the arkoses, and of the pelitic horizon, runs straight into the southern contact without contortion, and the dolerite appears to have made room for itself by stoping, or possibly by solution, of the arkoses from the sides of the plug, rather than by forceful intrusion. The complete absence of xenoliths of Moine rocks in the plug suggests that the dolerite was flowing over a considerable period of time, perhaps as a feeder to a lava flow, and this is confirmed by the wide aureole of thermal metamorphism surrounding the relatively small plug. There is no sign of any strongly chilled margin to the dolerite, and although the grain size is smaller in the outer 3 ft. of the plug than in the centre, the dolerite at the contact with the arkoses is still relatively coarse-grained.

Petrography.

Dolerite. Modal analyses and details of the optical properties of the principal minerals of the dolerite are given in table I.

The central part of the plug consists of olivine-dolerite with subophitic to intergranular texture in which anhedral olivine and clinopyroxene crystals up to 0·5 mm. across show a tendency to clot together in aggregates of grains. The average length of the plagioclase laths is 1·0 mm., but occasional crystals are up to 3·0 mm. long. Carbonate and a mesostasis of greenish chloritic material, which includes fine needles of pyroxene and apatite, occur interstitially to the plagioclase laths. Apart from partial serpentinisation of the olivine, the rock is very fresh.

About 3 ft. from the contact, the dolerite becomes slightly finer-grained, the average length of the plagioclase laths being 0·7 mm., with rare crystals up to 2·0 mm. long. The texture of the rock is subophitic. Olivine is absent, but may possibly be represented by patches of serpentine and carbonate. Spherical amygdales up to 2·0 mm. across are prominent in the outer part of the plug, and may be filled with carbonate or, rarely, with a zeolite mineral (not identified); carbonate is also common as grains interstitial to the plagioclase. The mesostasis consists of yellow isotropic material with pyroxene needles, and sometimes contains small isolated grains and aggregates of quartz.

At the contact with the Moine arkoses, the dolerite has much the same

TABLE I. Modal compositions (volume %) of some dolerites from the Glenmore plug, and optical properties of the principal minerals.

	1	2	3	4
Olivine ...	6·7	—	—	—
Pyroxene ...	27·3	23·0	25·0	25·4
Plagioclase ...	59·6	56·1	62·0	58·0
Iron ore ...	4·3	4·8	4·3	2·4
Carbonate ...	0·5	5·0	—	1·6
Mesostasis ...	1·6	11·1	—	12·6
Limonite ...	—	—	8·7	—
Olivine				
β ...	1·750			
2V ...	74°			
Pyroxene				
β ...	1·691	1·692	1·689	1·687
2V ...	47°	50°	49°	50°
Plagioclase				
Composition*	An_{71}	An_{72}	An_{73}	An_{71}

* Determined from the maximum extinction angle in the zone normal to (010).

1. Dolerite at the centre of the plug.
2. Dolerite 3 ft. from the contact.
3. Dolerite in contact with the fused arkose.
4. Dolerite in the apophysis along the line of the thin pelitic band.

appearance as elsewhere in the outermost 3 ft. of the plug, but contains a large proportion of limonite, possibly a weathering product of the mesostasis, which is not otherwise visible in this part of the plug. In some rocks, however, there is a patchy distribution of felsitic material, quite unlike the mesostasis of the normal types of dolerite in the plug, but very similar to the product of devitrification of the glassy part of the fused Moine arkoses, which will be described later. This felsitic material is partly interstitial and partly forms irregular rounded areas, suggesting that it may represent small fragments of arkose in various stages of assimilation into the dolerite.

In the apophysis along the line of the pelitic band, the dolerite shows a well-developed ophitic texture, due to the increased size of the pyroxene crystals, which are here up to 2·0 mm. across. The plagioclase laths are usually 0·5 mm. long, with rare crystals up to 1·5 mm. long. Spherical amygdales, usually 1·0 mm. in diameter, but sometimes as much as 7·5 mm. across, are common and are filled with carbonate. The pyroxene and plagioclase of the dolerite are both fresh, but the abundance of dark green chloritic mesostasis gives an altered appearance to the rock.

The field relations of the plug give no information regarding its age, except that it is later than the metamorphism of the Moine Schists. The freshness of the dolerite and the presence of olivine suggest that it is more likely to belong to the Tertiary igneous phase than to the Permo-Carboniferous suite of quartz-dolerite dykes and bosses.

Moine arkoses and pelite. The rocks into which the dolerite plug is intruded are thick, massive, false-bedded and pebbly, grey or white arkoses and belong to the part of the Moine Series referred to as the Glenmore Psammitic group by Richey *et al.* (1938, p. 71). The arkoses are composed of quartz, microcline, and plagioclase, variable amounts of muscovite and biotite, and accessory epidote, garnet, sphene, zircon, iron ore, apatite, and carbonate. Pelitic bands are rare in the Glenmore Psammitic group, but one is exposed *in situ* and in some large boulders about 45 ft. to the south-east of the plug; it is 4 in. thick, and is associated with micaceous and pebbly arkoses, with a subsidiary $\frac{1}{2}$-in. thick seam of similar pelite on the north-east side of the main band; similar pebbly arkoses and a thin subsidiary pelite are associated with the pelitic band metamorphosed at the contact, and this almost certainly represents the identical horizon within the arkoses. A chemical analysis of the pelite is given in table IV, together with the modal composition and optical properties of the principal minerals.

Metamorphosed arkoses and pelite. Due to the thick soil and detritus cover surrounding the dolerite plug, the metamorphosed Moine arkoses are only sporadically exposed. However, a nearly complete section through the metamorphic aureole of the plug was obtained by digging a trench along the line of strike of the pelitic band on the south-east side of the plug, and specimens collected from the rock *in situ* in the trench enable the sequence of changes with progressive metamorphism to be followed in both the arkose and the pelite from about 4 ft. up to a few inches from the contact with the dolerite.

The thermally metamorphosed arkoses are grey, black, or greenish-black tough rocks with a brittle fracture and vitreous appearance in which relics of the original pebbly texture can sometimes be distinguished. The pelites are tough, splintery, grey or black rocks with pale green streaks and bands parallel to the original schistosity planes.

In the specimens farthest from the contact, the micas of both the pelite and the arkose are already thoroughly altered. The original muscovites are turbid yellow in colour, and the biotites are almost completely opaque due to the formation of iron ore, but with traces of the original brown mica still showing the pleochroism and high bire-

fringence. A thin rim of finely crystalline potash feldspar is sometimes present as a reaction product between both types of mica and adjacent quartz grains. Garnet is still present in spongy grains, as in the original pelite, but is associated with iron ore, which appears to have separated from it. The arkose is shattered and some of the larger feldspar crystals are granulated, but there is no sign of fusion at this stage. The cross-hatched twinning of microcline is still present in the potash feldspar, which is biaxial with a high optic axial angle.

Three feet from the contact, the pelite presents much the same appearance as before, but the rims of potash feldspar between the original mica flakes and the quartz grains are more distinct (see pl. XIX, fig. 3). Basal sections of the original muscovite flakes show a hexagonally-arranged pattern of fine needles, probably of mullite (pl. XIX, fig. 4), but it was not found possible to separate the needles for precise identification on account of their very small size. In the arkose, the margins between the quartz and the feldspar grains are diffuse, and in some cases a very thin rim of glass can be distinguished. The potash feldspar is now sanidine with a small optic axial angle, and the reaction rims round the micas of the pelite can also now be identified as sanidine.

Two and a half feet from the contact with the dolerite, the potash feldspar rims (largely sericitized) round the original mica flakes of the pelite are 0·025 mm. wide, and all the relict mica has disappeared. A distinct rim of glass is always found between the quartz and the feldspar of the arkose. Relict garnet is still found in the pelite, but most of it is completely altered to iron ore and fine-grained turbid micaceous material, possibly an alteration product of cordierite.

Two feet from the contact, the reaction rims round the original micas of the pelite are up to 0·1 mm. wide and have a felsitic texture suggesting that they may represent devitrified glass from which some of the potash feldspar has crystallized as fine laths; greenish chloritic material occurs with the feldspar. The individual grains of opaque minerals derived from the biotite can now be resolved, and show the characteristic square and hexagonal cross-sections of spinel. The texture of the original schistosity of the pelite is still preserved up to this stage. In the arkose, parallel-sided channels of brownish glass are always present between the quartz and feldspar or altered mica. The feldspars have begun to recrystallize to a turbid, brownish mosaic, especially round the edges and along the cleavage cracks of the larger grains (pl. XIX, fig. 2). This resembles the initial stage of melting of albite observed experimentally by Day and Allen (1905, p. 120), a similar texture being described by

Guppy and Hawkes (1925, p. 329) as 'finger-print structure', and used by them as evidence of incipient melting of soda-orthoclase xenocrysts in a dolerite dyke in Iceland. The same texture in feldspar xenocrysts in Permian lavas in Devon was referred to as 'frit-lattice' or 'frit-channelling' by Tidmarsh (1932, p. 733), who attributed it to rapid heating and disintegration of the feldspar as a result of thermal expansion, the glass being considered to be a product of diffusion of the magma into the channels thus formed. The texture is also described by Lacroix (1893), Richarz (1924), Holmes and Harwood (1932), Davidson (1935), Knopf (1938), Larsen and Switzer (1939), Skelhorn (in discussion of Ackermann and Walker, 1960), and Wyllie (1961); the majority of these authors ascribe the texture to incipient melting of the feldspar. The small crystals of new feldspar preserve the optical orientation of the old, and, in the aureole of the Glenmore plug, as the contact with the dolerite is approached the finger-print texture spreads to cover the whole of the original detrital grain, so that eventually none of the clear feldspar remains, but the mosaic of small crystals still behaves optically as one unit. The margins of each unit frequently consist of slightly coarser crystals, suggesting that there has been some growth of new feldspar (cf. Larsen and Switzer, 1939, fig. 6, and Wyllie, 1961, p. 9).

Fifteen inches from the contact with the dolerite, the original schistosity of the pelite is locally destroyed, but trains of opaque spinel grains up to 0·03 mm. across indicate the positions of the original biotite flakes. Potash feldspar is abundant in the pelite as fine laths and skeletal crystals set in a turbid, brownish, devitrified glass, which contains rare spherical amygdales filled with carbonate, of very similar appearance to those found in the dolerite in the outermost zone of the plug.

Specimens of arkose and pelite close to the contact show the development of abundant glass, now usually turbid and brown in colour due to devitrification. In the pelite, spinel is very common in euhedral grains up to 0·1 mm. across; the smaller grains are translucent and dark brown. Thin needles of hypersthene up to 1·0 mm. long are found in both the pelite and the arkose, and feldspar occurs as lath-shaped skeletal crystals in the pelite, and as recrystallized mosaics in the arkose. Cordierite is rare, but occurs in some specimens in the characteristic rectangular and hexagonal cross-sections, often associated with recrystallized feldspar (cf. Ackermann and Walker, 1960, p. 244); the cruciform twinning of cordierite described by Wyllie (1959) in fused Torridonian arkose has not been found definitely in the aureole of the Glenmore dolerite plug.

Partly fused arkoses are found in many localities around the plug at

distances up to about 6 ft. from the dolerite. They are usually black-weathering, and on fresh surfaces present a black, pitchstone-like appearance. In thin section, the glass of these buchites is almost always altered, with varying degrees of devitrification from uniformly- or patchily-brownish isotropic glass to complete recrystallization, with the occasional development of microspherulitic texture. Embayed grains and aggregates of quartz and recrystallized feldspar represent the unfused portion of the original arkose. Microlites are very common in the glass of many of the buchites; the largest are prismatic crystals of strongly pleochroic hypersthene up to 1·5 mm. long, sometimes arranged in fan-shaped sprays of curved needles. Clinopyroxene is common, in the glass of some rocks, as fringes of short prismatic crystals around residual quartz grains. In the specimens of buchite from close to the contact with the dolerite, the margins of the quartz grains against the glass are slightly blurred through the formation of lath-shaped outgrowths, which in crossed polars extinguish in the same position as the neighbouring quartz grains. These probably represent tridymite that has subsequently inverted back to quartz, and are similar to the inverted tridymite fringes described and figured by Wyllie (1961, p. 16, and pl. 2, fig. B), who gives a bibliography of other occurrences of tridymite in fused siliceous rocks. Another type of microlite forms hair-like fringes surrounding the finger-print textured feldspars; these occur most commonly in the rocks in which the glass is partly devitrified, and in some cases it can be seen that the individual needles form prolongations from the edges of the coarser crystals at the margins of each unit of recrystallized feldspar. The remaining types of microlite are indeterminate due to their very small size; they take the form of fine opaque needles, sometimes with secondary growths resembling scopulites and margarites, and small euhedral opaque crystals, possibly of a spinel.

Veins. An interesting exposure of the contact between the dolerite and the buchite is found at the south-western corner of the plug (see fig. 1). A zone, 2 cm. wide, of white-weathering buchite occurs between the dolerite and the normal black, partly fused arkose. Contacts between this zone and the rocks on either side are sharp and thoroughly welded, but the boundary with the normal buchite has been accentuated by weathering whereas the contact with the dolerite is more difficult to see. A modal analysis of the dolerite is given in table I, which shows that it is of normal type but contains no olivine and has a large amount of limonitic alteration material. The light-weathering contact zone consists of embayed grains and aggregates of quartz and recrystallized feldspar in a

felsitic groundmass that contains sporadic clinopyroxene and twinned plagioclase crystals similar to those found in the dolerite. Quartz aggregates in this zone are frequently mantled by a halo of small clinopyroxene prisms. Locally this quartzose rock-type blends into patches of more

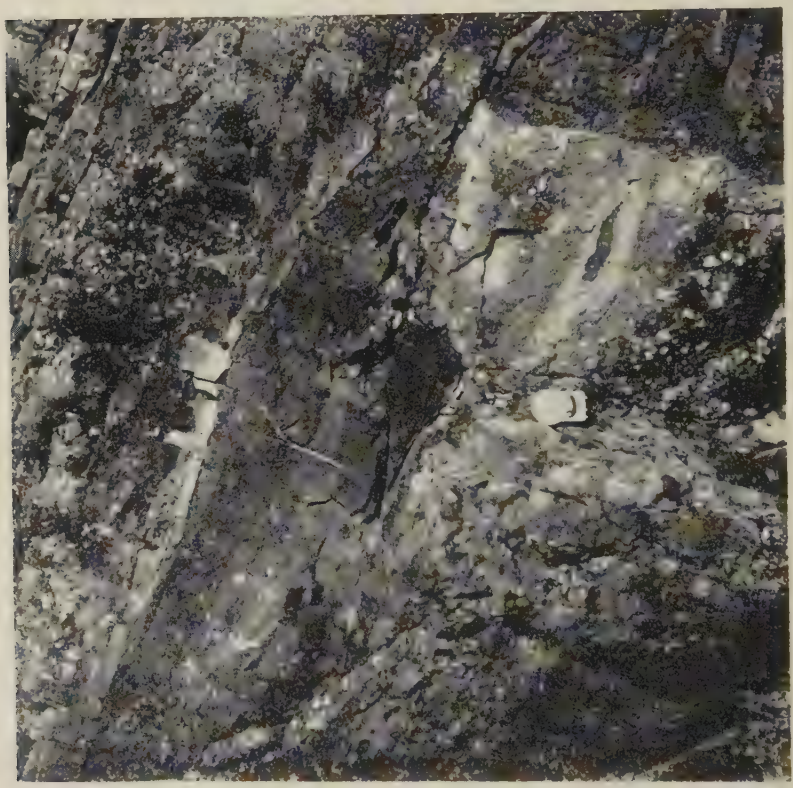

FIG. 1. Veining in arkoses at the contact on the south side of the Glenmore dolerite plug. The dolerite, showing good jointing, is on the left of the photograph. A zone of light-weathering arkose 2 cm. wide occurs between the dolerite and the black, smooth-weathering, fused arkose on the right, and from this zone a thin vein extends back into the fused arkose on the left of the exposure meter.

igneous appearance containing a larger proportion of pyroxene and plagioclase and a smaller amount of the fine-grained groundmass. A similar light-weathering zone of felsitic rock is found at most other exposures of the contact with the dolerite, though not usually so well displayed as here, and possibly represents a product of hybridism or contamination between the dolerite and the fused arkoses similar to that

described by Ackermann and Walker (1960) from the contact of the Heilbron sill. The sharp boundary between the modified contact zone and the normal buchite indicates a very abrupt limit of reaction, which in any case was of very small extent considering the much greater width of the aureole of metamorphism and fusion of the arkoses.

It appears to be from this contact zone that the veins shown in fig. 1 penetrate back into the normal buchites farther away from the contact. The veins are up to 4 mm. wide and the longest extends 20 cm. into the normal buchite, but since the dolerite contact is oblique to the rock-face the true penetration of the veins is somewhat less than this. The contacts of the veins with the buchite are sharp and the rock splits easily along the contact, which often contains chloritic and limonitic alteration. The veins are composed of a felsitic groundmass containing skeletal alkali feldspar and hypersthene needles (fig. 2); embayed quartz grains and aggregates are included in this groundmass, and so far as can be seen are never continuous with the similar quartz grains of the adjacent buchite. There is, however, no sign of flow textures or of preferred orientation of the included quartz grains in the veins. Except for the presence of the hypersthene and the absence of the igneous-looking clinopyroxene and plagioclase the material of the veins resembles the arkose of the contact zone. The central 0·4 mm. width of every vein consists of microcrystalline quartz forming a veinlet, usually straight but locally sinuous and bifurcating, which sometimes intersects and slightly displaces the embayed quartz grains included in the vein.

The form of the veins and the sharpness of their contacts with the normal buchite suggest that they may be rheomorphic, originating from the felsitic rock of the contact zone; on the other hand, their inhomogeneous composition and the presence of the unoriented quartz grains might indicate formation by metasomatism or simply by devitrification of the glass of the normal buchite, along fractures that were subsequently filled with the microcrystalline quartz. In view of the discontinuity of the quartz grains of the veins and of the buchite, it seems likely that some mobilization and movement of the material of the veins has occurred, and it is probably correct to class them as rheomorphic, even though they may have moved only a short distance, metasomatism and devitrification possibly having been the principal processes in their formation. A chemical analysis of one of the veins is given under C in table II.

A very clear small-scale example of rheomorphic veining was found, however, very close to the contact with the dolerite, in a specimen of

the thin pelitic horizon a few inches north-east of the main pelite described earlier. A vein of brown glass, 1·0 mm. wide, showing microspherulitic devitrification and containing green hypersthene microlites, intrudes

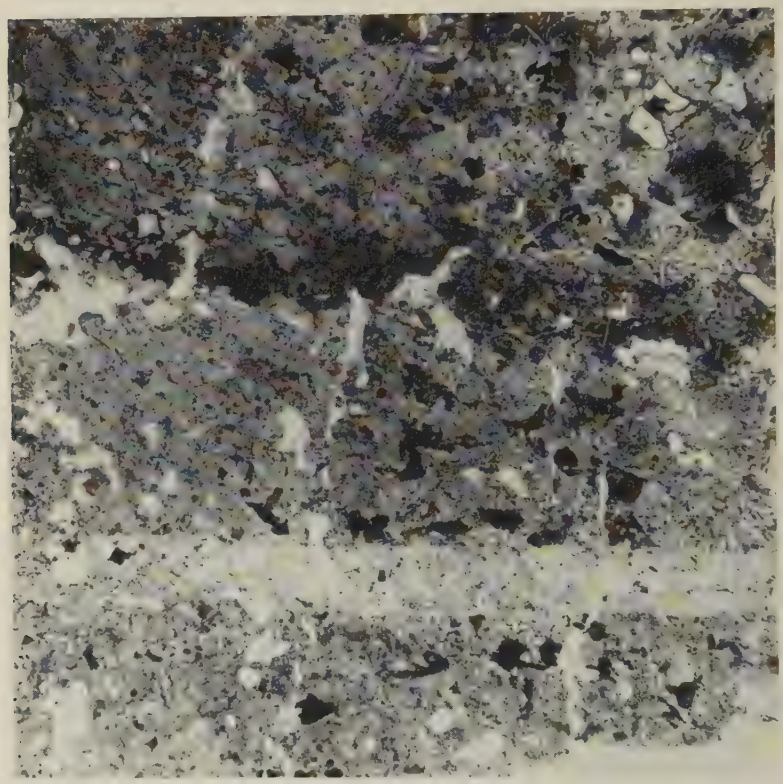

Fig. 2. A thin section of one of the veins shown in fig. 1. The vein, with felsitic texture, has a sharp contact against the glassy buchite at the top of the photograph, and the buchite is altered up to about 0·5 mm. from the contact. The veinlet of microcrystalline quartz in the centre of the vein cuts and displaces the two halves of a small embayed quartz grain just to the left of the middle of the lower half of the photograph. Ordinary light, ×16.

the pelitic band from the partly fused arkoses on either side (fig. 3). Residual quartz grains from the arkose have been strained off at the ends of the vein, but some of the smaller grains have been carried along with the liquid into the pelite. The rheomorphism displayed here is on a scale similar to the example described by Walker (1958) from the contact of a sill on the East Lomond, Fife.

Fig. 3. A thin pelitic band cut by a thermomorphic vein of devitrified glass from the adjacent partly fused arkoses. Large residual quartz grains are stranded off at the top end of the vein, and smaller grains carried into the pelite with the liquid in the vein. Ordinary light, ×6.

Yet another type of vein occurs in the analysed specimen, B, of the buchite 9 in. from the contact on the north-west side of the plug. The vein is 2·5 to 3·0 mm. wide; it has sharp contacts against the buchite and is clearly transgressive and of later date than the formation of the glass since it cuts across the embayed residual quartz and felspar grains

and the glass, which changes from colourless to brown within 0·5 mm. of the contacts of the vein. The vein consists of stumpy green clinopyroxene prisms 0·05 mm. long and skeletal alkali feldspar laths, sometimes showing a radiate arrangement, set in a deep brown glass with refractive index considerably less than that of canada balsam. This rock resembles the felsitic rock of the contact zone, and although the vein cannot be traced to a source in this zone it seems more than likely that it does represent mobilized material from this zone. The vein has not been included in the analysed portion of the rock.

Chemistry.

Two specimens of partly fused arkoses containing colourless glass of very fresh appearance were selected for chemical analysis. Both

TABLE II. Chemical analyses of thermally metamorphosed arkoses and their glasses, and a 'rheomorphic' vein.

	A, 3 ft. from contact.		B, 9 in. from contact.		C, Vein.
	Arkose.	Glass.	Arkose.	Glass.	
SiO_2 ...	70·15	70·42	74·44	71·60	73·12
TiO_2 ...	0·75	0·28	0·35	0·32	0·20
Al_2O_3 ...	13·24	12·54	11·71	12·14	10·66
Fe_2O_3 ...	1·90	0·89	1·14	0·79	4·34
FeO ...	1·43	0·67	0·85	0·80	0·68
MnO ...	0·03	0·03	0·02	0·02	tr.
MgO ...	1·08	0·52	0·75	0·62	0·88
CaO ...	2·58	0·96	1·98	0·79	1·92
Na_2O ...	2·88	3·29	2·97	3·68	2·59
K_2O ...	2·25	3·60	1·85	2·97	3·47
P_2O_5 ...	0·05	0·04	tr.	tr.	0·06
CO_2 ...	tr.	n.d.	0·03	0·00	tr.
H_2O^+ ...	3·03	6·43	3·86	6·07	1·54
H_2O^- ...	0·84	0·77	0·48	0·36	0·96
Total ...	100·21	100·44	100·43	100·16	100·42

New chemical analyses by B. C. M. Butler.

Modal composition of the arkoses

	A.	B.
Glass ...	40·7	55·2
Crystallites ...	2·6	1·0
Quartz ...	20·8	20·1
Feldspar ...	22·5	22·0
Opaques (mainly after mica)	13·4	1·7
Calcite ...	<0·1	—
	100·0	100·0
Refractive index of the glass (±·002)	1·496–1·500	1·499–1·502

specimens are from the north-west side of the plug; A was collected 3 ft., and B 9 in., from the contact. A sample of the glass was obtained from each rock by centrifugal separation with methylene iodide diluted with carbon tetrachloride; the glasses could not be obtained completely free from microlites, which form, however, only a very small proportion of the purified material. The arkoses and the glasses separated from them were chemically analysed and the results are given in table II.

TABLE III. Chemical analyses of thermally metamorphosed arkoses and their glasses, and a 'rheomorphic' vein.

(Recalculated anhydrous)

	A, 3 ft. from contact.		B, 9 in. from contact.		C, Vein.
	Arkose.	Glass.	Arkose.	Glass.	
SiO_2 ...	72·82	75·53	77·46	76·40	74·68
TiO_2 ...	0·78	0·30	0·36	0·34	0·20
Al_2O_3 ...	13·74	13·45	12·19	12·95	10·89
Fe_2O_3 ...	1·97	0·95	1·19	0·84	4·43
FeO ...	1·48	0·72	0·89	0·85	0·69
MnO ...	0·03	0·03	0·02	0·02	tr.
MgO ...	1·12	0·56	0·78	0·66	0·90
CaO ...	2·68	1·03	2·06	0·84	1·96
Na_2O ...	2·99	3·53	3·09	3·93	2·65
K_2O ...	2·34	3·86	1·93	3·17	3·54
P_2O_5 ...	0·05	0·04	tr.	tr.	0·06
CO_2 ...	tr.	n.d.	0·03	0·00	n.d.
	100·00	100·00	100·00	100·00	100·00
Normative composition					
Q ...	39·12	37·26	46·54	38·46	40·81
or ...	13·89	22·78	11·39	18·73	20·88
ab ...	25·18	29·90	26·17	33·25	22·40
an ...	13·33	5·01	10·21	4·18	7·38
C ...	1·43	1·63	1·28	1·52	—
en ...	2·81	1·41	1·94	1·65	1·52
of ...	—	—	0·05	0·29	—
di ...	—	—	—	—	1·53
mt ...	2·55	1·39	1·74	1·23	1·65
il ...	1·52	0·61	0·68	0·65	0·39
hm ...	0·16	—	—	—	3·30
ap ...	—	—	—	—	0·13
	99·99	99·99	100·00	99·96	99·99
Q+or+ab	78·19	89·94	84·10	90·44	84·09

A sample of the 'rheomorphic' vein shown in fig. 1 was obtained by cutting thin slabs from the unweathered interior of the rock and chipping the vein free from the adjacent buchite; the vein broke away easily along the contact with the buchite and thus it was easy to obtain a pure

sample. The chemical analysis of the vein is given under C in table II. All the chemical analyses have been recalculated anhydrous, and are given, with the normative compositions, in table III.

The normative quartz and alkali feldspars of the buchites (A and B), the glasses, and the rheomorphic vein (C) are plotted on the triangular diagram $NaAlSi_3O_8$–$KAlSi_3O_8$–SiO_2 in fig. 4, with additional compara-

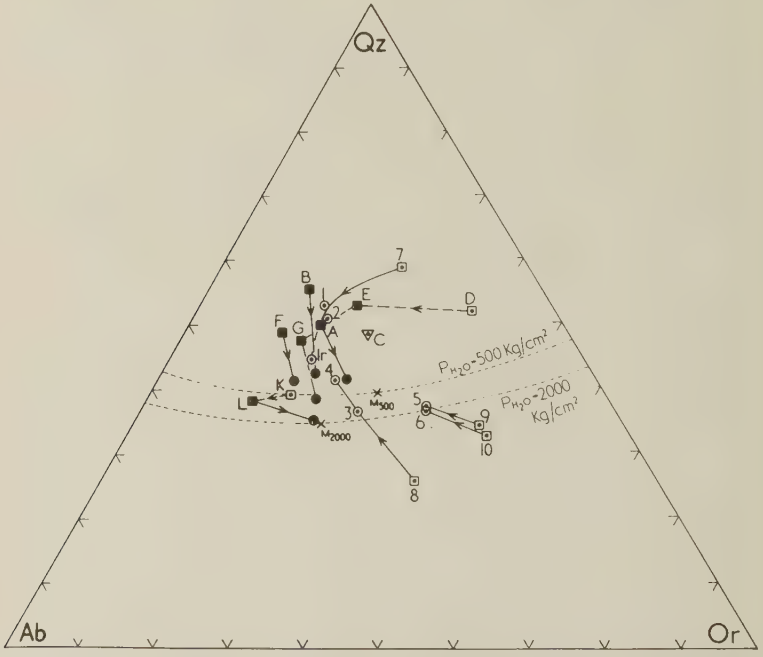

Fig. 4. Normative compositions of buchites, glasses, and associated rocks, plotted on the phase diagram $NaAlSi_3O_8$-$KAlSi_3O_8$–SiO_2. ⊡ unaltered parent rock; ■ buchite, connected by tie-line to ● glass separated from buchite; ⊙ palingenic dykes in the Satakunta diabase (Kahma, 1951); ▽ 'rheomorphic' vein in buchite of the Glenmore plug. The letters and numbers are explained in the text.

tive data. The cotectic boundary curves between the quartz and the feldspar fields are shown for pressures of water vapour of 500 and 2 000 Kg./cm². (Tuttle and Bowen, 1958, figs. 22 and 24), and the positions of the ternary minima on these curves are shown as M_{500} and $M_{2\,000}$ (Tuttle and Bowen, 1958, fig. 38). The following additional points are plotted for comparison with the data for the Glenmore plug:

F and G represent two pairs of buchites and their glasses, 18 in. and

1 in. respectively away from the upper contact of a sill of Karroo dolerite intruded into arkose of the Middle Ecca Series (Permian) near Heilbron, Orange Free State; D represents the composition of the unmetamorphosed arkose, and E that of a buchite 4 ft. from the contact (Ackermann and Walker, 1960).

K represents the composition of the Green Valley tonalite (part of the Southern California batholith): L is the same rock, metamorphosed and partly fused, forming a large inclusion in a Tertiary andesite plug, and the composition of the glass separated from L is shown by the solid circle close to the position of the ternary minimum $M_{2\,000}$ (Larsen and Switzer, 1939).

Points 1 to 10 and 1r represent the compositions of palingenic dykes and their parent rocks (granites and sandstone) associated with the pre-Cambrian Satakunta diabase in south-western Finland (Kahma, 1951, p. 32 and fig. 7 —the analyses are numbered as in Kahma's paper).

The Glenmore analyses and tie-lines are very similar in position and trend to the data from the Heilbron sill, and indicate a similar mode of formation. The composition of each glass, relative to the parent rock, lies closer to the ternary minimum of the silica–alkali-feldspar system, and the formation of the glasses by fractional fusion of the lowest-melting composition from the parent rock is thus clearly indicated. The relative positions of the two Glenmore glasses in relation to the ternary minimum are as would be predicted from field and petrographic evidence: B comes from a locality closer to the contact with the dolerite than A, and the higher temperature of metamorphism of B is shown both by the higher modal percentage of glass in the rock (table II), and by the greater departure of the composition of the glass from the composition of the ternary minimum.

By contrast with the close approach of the glassy products of partial fusion of rocks to the low-temperature trough of the 'residual system of petrogenesis', the composition of the vein C lies at a considerably greater distance from the position of the ternary minimum. It was shown from the petrographic evidence (p. 875) that the vein might be either rheomorphic, derived from the contact zone of the buchite, or formed *in situ* by devitrification of the glass of the normal buchite, with or without addition of other constituents. In either case, the chemical analysis, C, represents the bulk composition of a rock comprising the fine-grained groundmass, the residual quartz grains contained in it, and the central veinlet of microcrystalline quartz: the composition is therefore silica-rich relative to the felsitic material, which represents the

formerly liquid portion of the vein. The high Fe_2O_3 content of the analysis is probably due to the limonitic alteration along the contacts of the vein, which could not be excluded from the analysed material.

The palingenic dykes (nos. 3 to 6 in fig. 4) from granitic parent rocks associated with the Satakunta diabase lie close in composition to the ternary minimum of the Qz–Ab–Or system, indicating that there was partial or complete separation of the liquid from the partially re-fused granite. The palingenic dykes (nos. 1 and 2) from sandstone contain large rounded quartz and feldspar grains in the fine-grained micrographic groundmass, and these are interpreted by Kahma (1951, p. 32) as unmelted remnants of the original sandstone. The effect of these large grains was determined by calculation and subtracted from the rock composition to give the composition of the groundmass (1r), which represents the originally fused portion of the sandstone and lies close in composition to the ternary minimum.

Examples of rheomorphic veining have been described from the contacts of sills of Karroo dolerite, in association with which this phenomenon is particularly common (Walker and Poldervaart, 1949, p. 614). Chemical analyses of rheomorphic veins and mobilized sediments from sills of Karroo dolerite are given by Mountain (1944, 1960). The proportions of the normative Qz, Ab, and Or of these rocks appear to place them in two distinct groups; three of the analyses (Mountain, 1944, nos. I, II, III) fall within the silica field of the system Qz–Ab–Or, and may represent more or less unmodified sedimentary material, while four analyses (Mountain, 1944, no. IV, and 1960, nos. 1, 3, and 4) fall in or near the low-temperature trough extending from the position of the ternary minimum towards the eutectic point on the Ab–Or side of the triangle, and possibly represent sediments whose compositions have been altered by metasomatism. An analysis (Mountain, 1960, no. 2) of a granophyre associated with mobilized sediment no. 1 lies outside the trough on the Qz-Ab side. The compositions of these veins and mobilized sediments suggest that a variety of different processes may have been responsible for their formation.

Wager, Weedon, and Vincent (1953) show that the chemical composition of the Coire Uaigneich granophyre in Skye falls outside the range of normal granite compositions in the low-temperature trough of the Qz–Ab–Or system, and suggest that the granophyre may have been formed by the total fusion of quartz-rich material such as Torridonian arkose or acid Lewisian gneiss. If this is the correct inference to be drawn from the anomalous chemical composition of the granophyre, then it

provides an extreme example of the complete melting and mobilization of country rock resulting in the intrusion of a large body of magma.

It is interesting to compare these products of mobilization, or presumed mobilization, with rocks that are considered to be the products of metasomatic transfusion associated with Karroo dolerite sills. Whereas the rheomorphic veins and mobilized sediments, to which reference was made above, differ markedly in composition from the composition

TABLE IV. Chemical analyses of normal and thermally metamorphosed pelites.

	H.	I.	J.
SiO_2 ...	57·74	54·30	54·43
TiO_2 ...	1·25	1·37	1·36
Al_2O_3 ...	17·50	21·23	19·80
Fe_2O_3 ...	1·75	4·06	3·60
FeO ...	5·42	3·45	4·73
MnO ...	0·09	0·09	0·09
MgO ...	2·39	1·71	2·19
CaO ...	3·06	1·46	2·22
Na_2O ...	1·50	1·49	1·20
K_2O ...	5·86	7·28	6·93
H_2O^+ ...	2·36	2·78	2·50
H_2O^- ...	0·06	0·96	0·90
P_2O_5 ...	0·22	0·21	0·19
CO_2 ...	1·27	0·07	0·13
Total ...	100·47	100·46	100·27

Modal analysis of H, and optical properties of the principal minerals.

	Vol. %
Muscovite (γ 1·610)	29·4
Biotite (γ 1·650)	24·2
Epidote (β 1·734)	1·9
Garnet (n 1·797)	0·3
Plagioclase (An_{26})	7·5
Microcline	3·0
Quartz	30·4
Sphene	0·7
Carbonate	2·3
Apatite	0·3
	100·0

H. Normal Moine pelite 45 ft. south-east of the Glenmore plug.
I. Thermally metamorphosed pelite 3 ft. from the contact.
J. Thermally metamorphosed pelite 15 in. from the contact.
New chemical analyses by B. C. M. Butler.

of the ternary minimum in the Qz–Ab–Or system, five out of seven analyses of metasomatic granophyre produced by transfusion *in situ* of Karroo sediments (Walker and Poldervaart, 1949, Anals. 90, 92, 94, 95, 96) fall on the quartz–feldspar cotectic boundary line close to the position of the ternary minimum, even though the total Qz + Ab + Or is in no case greater than 80 % of the normative composition. These relations suggest that solid-liquid equilibria as well as metasomatism may have had some control over the compositions of the products of transfusion.

Three specimens of the pelitic band metamorphosed by the Glenmore plug were chemically analysed; the results are given in table IV. The

principal differences in composition between the normal and the thermally metamorphosed pelite are an increase in alumina and potash contents and in the Fe_2O_3: FeO ratio, and a decrease in silica, lime, and carbon dioxide; soda is nearly constant in amount, but possibly shows a slight decrease in the specimen closest to the dolerite contact.

Variation in alkali contents of thermally metamorphosed rocks. As mentioned earlier, the dolerite plug and the immediately adjacent metamorphosed Moine arkoses are surrounded on all sides by detritus, and consequently it is not possible to trace any individual band of fresh buchite from the contact with the dolerite into its unmetamorphosed equivalent farther away. (There is no doubt about the correlation of the pelitic horizon within and outside the metamorphic aureole, but unfortunately the glass of the arkose buchites close to the pelitic band is always altered.) In order to examine any possible changes in composition as a result of the contact alteration by the dolerite, a number of specimens of various types of normal Moine arkose were collected from exposures *in situ* within 100 yds. along the strike of the Moines to the south-east of the plug, and the alkali contents of five of these specimens were determined. The results of the analyses are shown in table V, with some comparative data.

The value of the molecular ratio Na/K is nearly constant in the five arkoses that have not been thermally metamorphosed (the average value for which is 1·11). There is a marked progressive increase in the Na/K ratio in the metamorphosed arkoses (reaching a value of 2·44 in the buchite 9 in. from the contact), which is due both to an increase of Na_2O and to a decrease of K_2O as the contact with the dolerite is approached. By contrast, the thermally metamorphosed pelites show a decrease in Na/K ratio compared with the same pelitic band outside the metamorphic aureole, and this decrease is due to an increase of K_2O. The variation of potash content of the pelites is rather larger than would be expected as a result of sedimentary variation over such a small distance, and the content of 7 % of potash in the thermally metamorphosed pelites is considerably higher than that of any other pelite in a series of twenty-two typical Moine schists that have been analysed from the Ardnamurchan area (B. C. M. Butler, unpublished data), the highest value for which was 6·1 % of potash. The normal pelitic and semipelitic schists show a wide range in Na/K ratio from 0·035 to 0·87, and a comparison of this ratio for the metamorphosed pelites with an average value for the Moine schists is not therefore particularly significant because of the wide variation in chemical composition of the latter, and

because in the vicinity of the plug there are no other bands of pelitic rock which could be analysed for comparison.

From the data given in table V, it seems likely that the pelitic band

TABLE V. Alkali contents of some rocks from the vicinity of the Glenmore dolerite plug, and a comparison of certain quartzo-feldspathic rocks and their thermally metamorphosed equivalents.

		Weight %.		Molecular proportions.		Molecular ratio.
		Na_2O.	K_2O.	Na.	K.	Na/K.
Glenmore dolerite plug						
Dolerite at centre of plug		2·99	0·58	0·0482	0·0062	7·77
Dolerite 3 ft. from contact		2·68	0·55	0·0432	0·0058	7·45
Normal Moine arkoses: (1)		2·49	3·65	0·0402	0·0388	1·04
(2)		2·82	3·81	0·0455	0·0405	1·12
(3)		2·96	3·91	0·0478	0·0415	1·15
(4)		2·02	3·01	0·0326	0·0320	1·02
(5)		2·50	3·08	0·0403	0·0327	1·23
Average of normal Moine arkoses	Q.	2·56	3·49	0·0413	0·0370	1·11
Thermally metamorphosed arkoses (buchites)						
3 ft. from contact	A.	2·88	2·25	0·0465	0·0239	1·95
9 in. from contact	B.	2·97	1·85	0·0479	0·0196	2·44
'Rheomorphic' vein 4 in. from contact	C.	2·59	3·47	0·0418	0·0368	1·14
Buchite adjacent to vein		3·11	1·89	0·0502	0·0201	2·50
Normal Moine pelite	H.	1·50	5·86	0·0242	0·0623	0·39
Thermally metamorphosed pelites						
3 ft. from contact	I.	1·49	7·28	0·0240	0·0773	0·31
15 in. from contact	J.	1·20	6·93	0·0194	0·0736	0·26
Heilbron sill (Ackermann and Walker, 1960)						
Arkose 36 ft. from contact	D.	1·15	5·70	0·0186	0·0605	0·31
Buchite 4 ft. from contact	E.	2·75	3·20	0·0444	0·0340	1·31
Buchite 18 in. from contact	F.	3·75	1·85	0·0605	0·0196	3·09
Buchite 1 in. from contact	G.	3·50	2·30	0·0565	0·0244	2·32
Cerro de la Calaveras (Larsen and Switzer, 1939)						
Green Valley tonalite	K.	3·59	2·34	0·0579	0·0249	2·33
Partly fused tonalite in andesite	L.	3·92	1·69	0·0633	0·0179	3·54
Strontian (Guppy and Sabine, 1956)						
Tonalite	M.	4·48	2·87	0·0723	0·0305	2·37
Tonalite contact-altered by dolerite	N.	4·79	2·56	0·0773	0·0272	2·84
Ardmucknish (Flett, 1908)						
Phyllite 6 ft. from contact	O.	1·84	3·43	0·0297	0·0364	0·82
Buchite at contact with camptonite	P.	2·58	2·01	0·0416	0·0213	2·00
Soay (Wyllie, 1961)						
Average Torridonian arkose (from Kennedy, 1951)		2·45	3·35	0·0395	0·0356	1·11
Arkose xenolith in picrite sill		2·96	1·35	0·0478	0·0143	3·34
Flurrybridge (Reynolds, 1937)						
Newry granodiorite 15 yds. from contact with gabbro	R.	4·09	2·95	0·0660	0·0313	2·11
Granodiorite 10 yds. from contact	S.	3·90	3·82	0·0629	0·0406	1·55
Waterfoot (Reynolds, 1940)						
Triassic sandstone 9 ft. from contact with dolerite dyke		0·84	1·74	0·0136	0·0185	0·74
Sandstone 4 in. from contact		0·79	2·52	0·0127	0·0268	0·47
Mullaghbawn (Reynolds, 1941)						
Newry granodiorite 25 ft. from contact with gabbro	T.	4·41	2·95	0·0712	0·0313	2·27
Granodiorite 11 ft. from contact		3·22	3·80	0·0520	0·0403	1·29
Granodiorite 8 ft. from contact	U.	3·34	4·28	0·0539	0·0454	1·19
Granodiorite 2 ft. from contact		3·40	3·06	0·0549	0·0325	1·69
Granodiorite 2–3 in. from contact	V.	4·76	2·01	0·0768	0·0213	3·61
Langesundsfjord (Goldschmidt, 1914)						
Normal metamorphosed sandstone*	W.	0·89	1·79	0·0144	0·0190	0·76
Soda-rich metamorphosed sandstone*	X.	3·47	1·77	0·0560	0·0188	2·98
* CaO contents 4·86 % and 1·48 % respectively.						
Dinas Head (Agrell, 1939)						
Black carbonaceous slate (anal. 3)	Y.	0·51	4·57	0·0082	0·0485	0·17
Very fine-grained adinole (anal. 8)	Z.	8·89	0·31	0·1434	0·0033	43·5

gained potash, and the arkoses lost potash and gained soda in the metamorphic aureole of the plug. A possible explanation may be that there was an outward migration of potash from the buchites close to the contact, and that part of this potash was introduced into, and retained in, the adjacent pelitic band. The breakdown of muscovite and biotite in the metamorphosed pelite, and the formation of rims of potash feldspar between the micas and the quartz indicate reactions of the type:

$$\text{Muscovite} + \text{Quartz} = \text{Sanidine} + \text{Mullite} + H_2O,$$

$$\text{Aluminous biotite} + \text{Quartz} = \text{Sanidine} + \text{Spinel} + H_2O,$$

with a release of alumina (in the form of mullite and spinel) in excess of the amount required by the K_2O to form potash feldspar. Although the contents of both alumina and potash are higher in the metamorphosed than in the original pelite, there is a progressive decrease in the ratio Al_2O_3/K_2O (H, 2·99; I, 2·92; J, 2·86) which suggests that the excess of alumina was being used up in the pelites closer to the contact. This provides a possible mechanism for the fixing of part of the migrating potash in the pelitic band.

The alkali contents of the 'rheomorphic' vein, C, are very different from those of the normal buchite immediately adjacent to the vein, showing clearly that the vein cannot have been formed simply by devitrification of the glass of the normal buchite without change in chemical composition (cf. p. 875). The data do not, however, help to decide which of the other hypotheses of formation of the vein (by rheomorphism or by metasomatism along fractures) is likely to be correct. The alkali contents of the vein, C, are practically identical with the average alkali contents of the Moine arkoses, Q, outside the metamorphic aureole of the plug. In view of the marked changes of alkali content in the other rocks close to the contact with the dolerite, it is probable that this is a coincidence rather than that the vein represents the unmodified original composition of the arkoses at the contact, although it is hard to explain why the vein is potash-rich relative to both the adjacent buchites and the dolerite.

A comparison of the data for the Glenmore plug with other examples of metamorphism of quartzo-feldspathic rocks at the contacts with basic intrusions shows that movement of alkalis is very common or even ubiquitous in this type of environment. The data are summarized in table V and fig. 5. Instances of hybridization, granitization, and contact alteration in which there are important chemical changes in addition to variation in alkali contents have not been included, except for the special

case of alkali metasomatism that culminates in the formation of adinoles (Agrell, 1939). Only those analyses are plotted in fig. 5 for which there is convincing evidence of change in alkali contents.

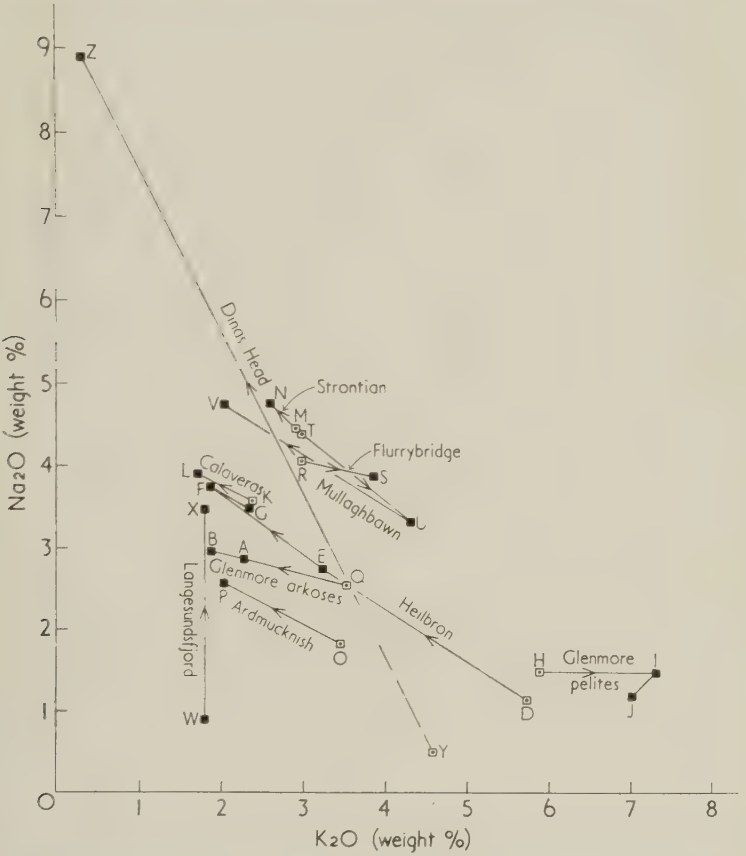

FIG. 5. Variations in alkali contents of quartzo-feldspathic rocks in thermal metamorphism. ⊡ unaltered parent rock; ■ thermally metamorphosed equivalent. The analyses are indicated by letters as in table V.

The metamorphism of arkose by the Heilbron sill (Ackermann and Walker, 1960) produces variations in the alkali contents analogous to those of the arkoses of the Glenmore plug, but with an even greater difference between the unmetamorphosed arkose and the buchite at the contact. A similar, but less extensive, variation is shown in the metamorphism of Green Valley tonalite by the Calaveras andesite (Larsen

and Switzer, 1939), and of Strontian tonalite by a quartz-dolerite plug (Guppy and Sabine, 1956, p. 11). These two occurrences are of particular interest since each tonalite could be analysed both in the original and in the metamorphosed state. Although the original tonalites may have been of variable composition, the fact that the trend of variation is essentially the same in each case makes it more likely that the changes in the alkali contents are the result of the metamorphism.

The chemical analyses of phyllite and of the same rock metamorphosed and partly fused by a camptonite dyke at Ardmucknish (Flett, 1908) show changes in alkali contents similar to those in the examples given above. Flett noted that the phyllite buchite was richer in soda and poorer in potash than the unmodified phyllite, but considered that 'there seems no reason to believe that there has been any transfer of alkalis' during the metamorphism. In view of the similarity of this occurrence to the others already quoted, it seems likely that the changes noted by Flett are significant, and that transfer of alkalis has in fact occurred. The high soda contents of buchite xenoliths from the composite sill at Carsaig in Mull (Thomas, 1922) also suggest similar metasomatic changes, but the compositions of the original shales are not known.

Wyllie (1961) remarks that the fused xenolith in the picrite sill in Soay has a distinctly lower K_2O/Na_2O ratio than the average of three Torridonian arkoses given by Kennedy (1951), but that the Soay Torridonian rocks are known to vary in composition. However, the low K_2O content of the xenolith relative to the average (see table V) suggests that here too there has been movement of alkalis resulting in a slight gain of soda and a marked decrease of potash in the metamorphosed arkose.

The analyses of the palingenic dykes from sandstone associated with the Satakunta diabase (Kahma, 1951) also indicate that some transfer of alkalis may have occurred in the formation of the dykes. In fig. 1 there is a close similarity in trend of the points 7–1–2 for the Satakunta sandstone and dykes and D E–F–G for the arkose and buchites from the Heilbron sill.

In every example quoted above there is a gain in soda and a loss in potash during metamorphism. It is worth noting that the decrease in the molecular proportion of the potash is not exactly balanced by the increase in the molecular proportion of the soda (i.e. there is not an ion-for-ion replacement), although in the metasomatism of the arkoses associated with the Heilbron sill this condition is nearly attained (cf.

table V, decrease in K from D to F − 0·0409, increase in Na + 0·0419). In all the examples fusion has occurred during the metamorphism, except in the case of the Strontian tonalite, although the separation of a lowest-melting composition represented by abundant micrographic intergrowths of quartz and alkali feldspar in the contact-altered rock (Guppy and Sabine, 1956, p. 11) may be the result of fractional melting. It appears probable that the alkali variation in these rocks is related to the formation of the glass.

The Glenmore arkoses are coarse-grained rocks containing two feldspars (microcline and sodic plagioclase) and mica. Although the plagioclase of the arkose will contain some potash feldspar in solid solution, and vice versa, fusion of such a rock would be expected to produce a mainly Qz–Ab liquid at quartz–plagioclase contacts and a mainly Qz–Or liquid at quartz–microcline or quartz–mica contacts. The liquid resulting from fusion of this kind would thus be inhomogeneous in composition. Yet staining of glassy Glenmore arkoses with sodium cobaltinitrite, using the method described by Chayes (1952), reveals no inhomogeneity of potash content of the glass, even though from the textural evidence (the parallel-sided channels of glass and the lack of alignment of microlites in them) it is clear that there has been virtually no movement of the glass or of the residual quartz and feldspar which could explain mechanical mixing. It is perhaps especially significant that the glass channels are of nearly constant width whether they are between quartz and feldspar or between quartz and relict mica (pl. XIX, fig. 2). One important effect of alkali metasomatism may thus be to homogenize the liquid once it is formed. But migration of chemical constituents may have an even more fundamental part to play in producing locally in the still solid rock a lowest-melting composition that is capable of being fused. A migration of soda to contacts of quartz with potash feldspar or mica or migration of potash to contacts of quartz with plagioclase will produce locally a composition much closer to the ternary minimum than that represented by the original phase boundaries, and it is possible that alkali migration of this kind is an essential precursor of fusion, without which melting may be difficult or impossible.

It would obviously be difficult to decide whether metasomatism preceded or followed fusion in any particular case, but two recently-described examples of the melting of rocks, one natural and the other experimental, are of particular interest. Ackermann and Walker (1960, p. 244) state that in the arkoses metamorphosed by the Heilbron sill 'the first glass is formed where grains of potash-feldspar and quartz

come into contact with one another'. The original arkose was potash-rich (point D in fig. 4); the first-melting liquid may therefore have had a composition on the quartz–feldspar cotectic, but potash-rich for lack of available soda and because of the relatively gentle gradient out of the ternary minimum towards the Qz–Or side (cf. Tuttle and Bowen, 1958, fig. 40), and the liquid subsequently gained soda by metasomatism to attain a composition closer to the ternary minimum. Alternatively the soda metasomatism that is so well displayed in these rocks may have enabled the fusion to take place by adding potential albite to the quartz–potash feldspar boundaries in the solid rock, and the first product of melting may have had a composition close to that of the ternary minimum.

Winkler (1957) investigated experimentally the metamorphism of illitic clays and was able to reproduce anatectic melting at 700° to 720° C. The starting material was potash-rich, and consequently the product of melting 40 to 65 % of the initial material was a potash-rich composition at the Qz–Or end of the quartz–feldspar cotectic boundary. However, the products of melting 23 to 34 % of illitic clay with varying contents of NaCl and $CaCO_3$ had compositions very close to the liquid produced by melting of the clay alone (Winkler and von Platen, 1958, table 3 and fig. 6; 1960, table 7 and fig. 1), and the liquid attained a more sodic composition only at a higher temperature when 39 to 66 % of the original material had melted. The results appear to indicate that the added NaCl had little effect on the composition of the earlier products of melting, but reacted with the liquid after it was formed so as to give a composition closer to the ternary minimum. This would correspond to the case of metasomatism following fusion of natural rocks.

The evidence from the Glenmore plug also has a bearing on this problem. If the first product of fusion of natural rocks is potash-rich, as might be suggested from the evidence of the two papers just referred to, then the earliest appearance of glass should be seen at the contacts of quartz with the potash feldspar reaction rims round the original micas of the pelitic band. In fact, the arkoses, which contain both potash feldspar and sodic plagioclase, are the first to show signs of fusion, while the potash feldspar rims in the pelite are still apparently crystalline; the potash-rich pelite only shows extensive melting in the specimens from closest to the dolerite contact. These results confirm that in the fusion of natural rocks the first melting occurs when soda and potash are present in approximately equivalent proportions, thus agreeing with the data from the simple experimental system $NaAlSi_3O_8$–$KAlSi_3O_8$–SiO_2, but

they do not show whether metasomatism precedes or follows fusion in this particular case.

The variations in alkali contents do not always directly aid the fusion, as is made clear in the example of melting of Green Valley tonalite (fig. 4). The effect of the metasomatism is to move the composition of the rock out of the low-temperature trough towards the Qz–Ab side, but the composition of the glass lies in the trough close to the ternary minimum for 2 000 Kg./cm.^{2} H_2O pressure. The opposite trends of the metasomatism and the fusion are well shown by the opposing directions of the tie-lines connecting K to L and L to its glass. The trend of metasomatism is probably related to the sodic environment of the xenolith in the andesite plug, a point which was noted by Reynolds (1941, p. 23 and fig. 6).

Examples of metamorphism in which there has been an increase of potash rather than soda have been described by Reynolds. At Flurrybridge in the Slieve Gullion area, Newry granodiorite has been metamorphosed by 'quartz-gabbro', and shows a marked increase in potash, while soda shows a slight decrease (Reynolds, 1937); the same relations are found in the metamorphism of a Triassic sandstone by a Tertiary dolerite dyke at Waterfoot, Co. Antrim (Reynolds, 1940), although in this case it is possible that sedimentary variation is the explanation for the differences between the unaltered and the metamorphosed sandstone. A more complex example of metamorphism of Newry granodiorite by a gabbro mass at Mullaghbawn (Reynolds, 1941) has resulted in a geochemical culmination of potash and depression of soda in the granodiorite at distances of 2 to 11 ft. from the contact, while at the contact itself there is a slight enrichment in soda and decrease in potash relative to the unmodified granodiorite. This is shown by the initial decrease and subsequent increase in the ratio Na/K as the contact is approached, and by the reversal of the trend of the tie-line connecting the points R, S, and T in fig. 5. The texture of the metamorphosed granodiorite figured by Reynolds (1937, plates 22 and 23; 1941, fig. 2 and plate 1) strongly resembles the parallel-sided glass channels produced by partial melting of igneous rocks (Knopf, 1938; Larsen and Switzer, 1939), but is ascribed by her to metasomatic replacement of the quartz grains of the granodiorite by felsitic material.

A more extreme example of metasomatism is the transfusion of quartz xenoliths in basic and ultrabasic potassic lavas described by Holmes (1936). In this case the added material entered the rock along sinuous channels of uniform width, usually following intergranular boundaries

but sometimes traversing individual grains. The appearance of the transfused rock is similar to that of the fused Glenmore arkoses, except that the channels of glass occur everywhere between quartz grains (see Holmes, 1936, pl. XIII, fig. 3), whereas at Glenmore, and in every other case of partial fusion to which reference has been made here, the glass is always developed between quartz and feldspar or relict mica, and never between grains of the same species. The occurrence is of interest in that the formation of the glass must be ascribed largely to the introduction of potash, alumina, water, and other common rock-forming oxides from the enclosing magma, only the silica being contributed by the xenolith itself; the composition of the glass lies on the Qz–Or side of the ternary minimum of the Qz–Ab–Or system (see Wyllie, 1961, fig. 5), probably due to the highly potassic environment of the xenolith during the metasomatism.

It is interesting to compare these instances of metamorphism of quartzo-feldspathic rocks by basic igneous rocks with the example of metamorphism and metasomatism of quartzite by nepheline-syenite pegmatite in Langesundsfjord described by Goldschmidt (1914). Here there is a marked increase in Na_2O in the metasomatized quartzite, but at the expense of CaO, which shows an equivalent decrease, while K_2O is unaffected in amount.

The metasomatism of the pelitic band at the contact of the Glenmore plug is of an entirely different type from the extreme soda metasomatism of shale at dolerite contacts that results in the formation of adinoles (Agrell, 1939; Davies, 1956), although the Glenmore arkoses and several of the other examples of metamorphism of quartzo-feldspathic rocks show a trend of soda enrichment similar to but much less marked than that of adinolization. The contrasted metasomatic effects produced in the arkose and in the pelite at Glenmore afford an interesting example of selective diffusion of soda and potash between the contrasted rock types.

Water contents of the buchites and glasses. The H_2O^- contents quoted in table II for the buchites and their glasses were determined from the weight loss after heating a 1-g. sample at 105° C. for 1 hour. The weight losses of two samples (A and the glass from A) were also determined by reweighing after heating for repeated 10-minute intervals. It was found that both samples lost nearly all their 'H_2O^-' in the first 10 minutes, and thereafter both samples showed a steady slow loss in weight, which was continued even after heating for continuous periods of 16 and 19 hours at the same temperature. The overall loss of weight after a total

of 42 hours' heating at 105° C. was equivalent to 1·3 % 'H_2O^-' in the buchite and 1·7 % in the glass; even after this time neither sample had reached constant weight. The results confirm the observations of Ross and Smith (1955, p. 1078) and Carmichael (1960, p. 326) that the loss of water is dependent more upon the duration than upon the temperature of heating.

The glasses of the buchites from Glenmore contain more than 6 % of H_2O^+ (H_2O retained after heating at 105° C. for 1 hour), which should correspond to a water vapour pressure, at the time of formation, of 2 000 Kg./cm.^{2} (Tuttle and Bowen, 1958, fig. 28), or a depth of about 7 km., which is almost certainly an over-estimate of the probable depth of cover at the time of intrusion of the plug. Comparable water contents (high H_2O^+ or H_2O^- or both) are found in many analyses of fused rocks and glasses (Flett, 1908; Thomas, 1922; Holmes, 1936; Larsen and Switzer, 1939; Frankel, 1950; Ackermann and Walker, 1960; Wyllie, 1961), and also in acid igneous glasses (Hawkes and Harwood, 1932; Tilley, 1957; Carmichael, 1960), and in residual glasses in tholeiites (Walker, 1935; Elliott, 1956).

Determination of the refractive-index:total-water-content relations of the glass of specimen A, following the method used by Ross and Smith (1955), gave the result:

Mean R.I. before heating (from table II)	$(n_h) = 1{\cdot}498$
Mean R.I. after heating	$(n_e) = 1{\cdot}483$
Total water content (from table II)	$(w) = 7{\cdot}20\%$

$$(n_h - n_e)/w = 0{\cdot}015/7{\cdot}20 = 0{\cdot}0021$$

This value of the ratio is considerably less than any quoted by Ross and Smith, and indicates clearly that the glass is of perlite, not obsidian type, and that the water content represents water of hydration introduced later than the formation of the glass (perhaps even as a result of weathering), and not the water content of the liquid portion of the buchite at the time of metamorphism. Consequently the water content of the glass cannot be used to determine the temperature and pressure of metamorphism.

The ease with which synthetic silicate glasses of 'granitic' composition can be crystallized in the presence of water vapour under pressure, compared with the difficulty of crystallizing the same glasses under anhydrous conditions (cf. Tuttle and Bowen, 1958, pp. 5–6), suggests that natural silicate melts with a high water content would crystallize rather than vitrify on cooling. The preservation of clear isotropic glass with high

water content in the Glenmore and other examples is therefore of interest. It is possible that the fused rocks originally had a low water content so that the conditions of solidification approached those of the anhydrous synthetic systems. More probably the vitrification is the result of quenching. The difference between the felsitic or granophyric texture and highly altered appearance of the metamorphosed arkoses of the majority of the dolerite plugs of Ardnamurchan and the fresh glassy buchites at Glenmore (cf. pl. XIX, figs. 1 and 2) is almost certainly related to the small size of the Glenmore plug. The wide metamorphic aureole indicates the continued flow of hot dolerite magma, but once the conduit became blocked, or the supply of magma ceased, cooling would be very fast due to the small heat content of the plug and the rapid dissipation of heat into the surrounding arkoses. The preservation of the glass and the absence of hydrothermal alteration in both the buchites and the dolerite is thus explained. The much larger size of the other dolerite plugs and dykes in Ardnamurchan allowed slower cooling, with crystallization of the liquid portion of the buchites and extensive alteration of the arkoses and the dolerite.

Conclusions.

The results of metamorphism of Moine arkoses by the Glenmore dolerite plug are essentially the same as those produced by fusion of Torridonian arkose in a picrite sill in Soay described by Wyllie (1961), who gives a full discussion of the conditions of metamorphism. Since the Glenmore rocks do not add any further information bearing on the problem of the temperature and pressure of metamorphism, it is not necessary to review all the data again here. It may be remarked that the presence of tridymite fringes round quartz grains in the buchites close to the dolerite contact permits only a restricted range of possible temperatures and pressures because of the small stability field of tridymite (Tuttle and England, 1955), provided it can be assumed that the tridymite was not formed metastably (cf. Wyllie, 1961, p. 24).

The wide metamorphic aureole around the plug and the absence of arkose xenoliths within it suggest that the dolerite was flowing over a considerable period of time. The lack of contortion of the quartzites, either on a large or on a small scale, is remarkable, considering that they were more than half liquefied; it seems probable that the liquid portion of the buchites was very viscous, and this would also explain the limited extent of the zone of hybridization, which is only 2 cm. wide. It is possible, however, that some reaction of the dolerite with the arkose

took place, since olivine is apparently absent and small ovoid patches of felsitic material are present in the outer part of the plug; such reaction probably took place by assimilation of arkose, either stoped or dissolved from the sides of the plug. Solution of arkose can have had only a very small effect on the bulk composition of the dolerite, and the total volume of dolerite that flowed up the conduit must have been considerably greater than the volume of arkose that had to be removed to make room for it.

The product of partial fusion of the arkose is a buchite composed of 40 to 55 % of a glass with a composition close to the ternary minimum in the synthetic system $NaAlSi_3O_8$–$KAlSi_3O_8$–SiO_2, and the formation of the glass by fusion of the lowest-melting composition from the arkose is thus clearly indicated. In addition to the fusion, the arkoses show migration of alkalis resulting in a decrease of K_2O and an increase of Na_2O in the buchites close to the contact with the dolerite, while an associated pelitic band shows an increase of K_2O and a slight decrease of Na_2O. A review of other examples of thermal metamorphism of quartzo-feldspathic rocks by intrusions of basic magma shows that alkali metasomatism is a very common feature of this kind of occurrence, and it is suggested that local alkali migration is related to the fusion process.

Acknowledgements. My thanks are due to Prof. C. E. Tilley, F.R.S., for his advice and encouragement during this investigation, and to Prof. Tilley and Dr. G. M. Brown for critically reading the manuscript. The work was carried out partly during the tenure of a D.S.I.R. Research Studentship at the Dept. of Mineralogy and Petrology, Cambridge University, and partly at the Dept. of Geology and Mineralogy, Oxford University.

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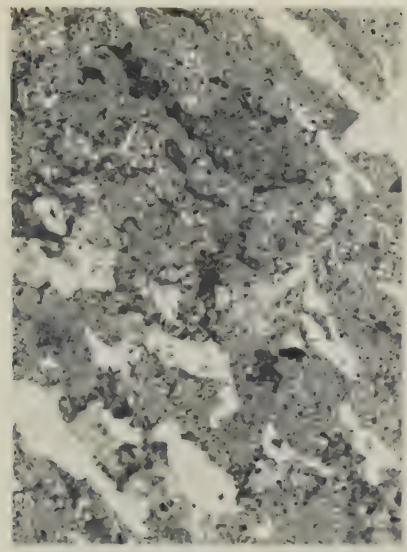
Explanation of Plate XIX.

Fig. 1. Thermally metamorphosed Moine arkose close to the contact on the south side of the dolerite plug at Camasinas, Ardnamurchan (Grid reference 656608). Embayed residual quartz grains are enclosed in a granophyric groundmass containing carbonate, apatite, and iron ore. The lath-shaped crystals in the groundmass extinguish in the same position as the adjacent quartz grains and possibly represent inverted tridymite. Magnification ×17. The rock is very similar in appearance to the metamorphosed Torridonian arkose figured by Wager, Weedon, and Vincent (1953, pl. XIII, fig. D).

Fig. 2. Partly fused Moine arkose 3 ft. from the contact of the Glenmore dolerite plug (analysed specimen A, table II). Parallel-sided channels of glass (mottled appearance) occur between quartz (white) and feldspar (grey) or original mica flakes (black). A feldspar grain partly recrystallized along cleavage cracks can be seen at the centre of the photograph. Magnification ×30.



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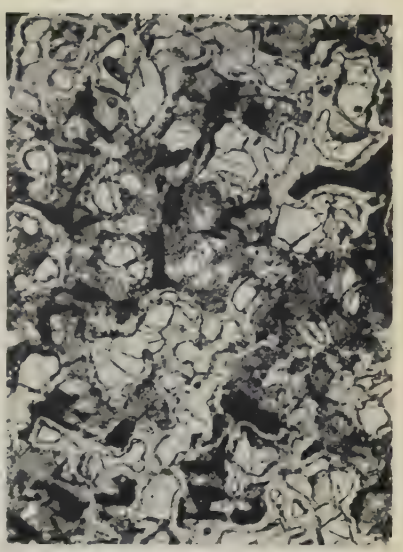

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P. C. M. Butler: Metamorphism and metasomatism

FIG. 3. Thermally metamorphosed pelite (analysed specimen I, table IV). The relict schistosity can still be seen, but the micas are almost completely altered and reaction rims of potash feldspar occur between them and the quartz grains. Magnification ×30.

FIG. 4. A basal section of original muscovite in a thermally metamorphosed pelite 3 ft. from the contact with the dolerite. The fine needles are probably of mullite, in hexagonal arrangement relative to the original muscovite crystal. Magnification ×230.

All photographs in ordinary light.

Electron-optical and electron-diffraction study of a disordered structural state in the transformation goethite–hematite.

(With Plate XX.)

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and

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[Read 23 March 1961.]

Summary. Previous X-ray observations by J. Lima-de-Faria and P. Gay on the transformation goethite–hematite indicate that in the process of state change at low temperatures, maxima of scattered X-ray intensity occur as satellites to strong and diffuse Bragg maxima due to hematite. These maxima define a repeat distance of about 32 Å. Corresponding satellite maxima have been observed using electron diffraction technique and satellite maxima flanking the direct beam (000) have been used to obtain a direct electron optical image of the disordered state at high magnification. Such photographs show a mainly regular fringe system with spacing of approximately 30 Å. Dislocation patterns were also observed.

THE present paper contains an account of the direct electron optical resolution of a characteristic repeat in a transitional structural state in the transformation goethite–hematite. The observations on which this paper is based were made subsequent to a detailed study of this transformation by single-crystal X-ray technique (Lima-de-Faria and Gay, in press).

During the latter investigation strong maxima of scattered X-ray intensity were observed as satellites to Bragg maxima due to hematite. Both the intensity of these satellite maxima, and the scale of the periodicity which they defined, about 32 Å., were considered favourable to possible direct-electron optical resolution of the disordered state.

The instrument used for the present experiments was a Siemens Elmskop I, which was operated throughout at 100 kV. A selected-area diffraction aperture defining a specimen area of diameter 0·25 μ was used in the associated diffraction experiments.

The material used in the present study was a single crystal of goethite,

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which was previously heated at 300° C. for 19 hours in air and subsequently examined by single-crystal X-ray oscillation technique. The heat-treated material was then crushed and mounted on a carbon film using a standard electron microscope supporting grid.

On diffraction, fragments of this material were found to yield reciprocal-lattice sections indicating that the crushed fragments had a tendency to lie either on (0001) or $(10\bar{1}0)$.[1]

A detailed selected-area diffraction study (with 0·25 μ aperture) was made on a flake lying on $(10\bar{1}0)$ (i.e. with a and c^* in the plane of the flake). All the selected area diffraction photographs from this flake showed consistent development of satellite maxima along c^* rows through the Bragg maxima. On the original plates weak second order maxima of intensity were also observed. One of the diffraction photographs in this series has been illustrated and indexed in pl. XX, fig. A, and in fig. B the position of the selected area which provided this diffraction pattern has been indicated.

The nature of the off-on high voltage fluctuations on the microscope used does not permit of an accurate determination of absolute values for the cell-dimensions, but, because the reciprocal lattice plots obtained comprise very sharp diffraction maxima and because these plots are effectively undistorted, it has been possible to derive accurate values for the ratios of the reciprocal distances concerned by direct measurement. The c/a ratio of hematite determined in this way was $2{\cdot}76 \pm 0{\cdot}01$ and the ratio of the repeat defined by the first-order satellite maxima with respect to the c^* repeat of hematite was determined in the same way as $2{\cdot}50 \pm 0{\cdot}02$, corresponding to a true repeat in direct space of 34·4 Å. in terms of the c-dimension of hematite (13·75 Å.). This repeat is equal to approximately 7·5 times the a repeat of goethite. The $(10\bar{1}0)$ flake initially used for detailed selected-area diffraction study was also used to obtain a direct electron-optical image at high magnification ($\times 80\,000$). This flake has been illustrated in pl. XX, fig. B. In this case a suitable objective aperture was used to select the direct beam (0000) and its associated satellite maxima for image formation. The resulting photograph showed a well-developed fringe system with spacing of approximately 30 Å. An enlargement of part of this photograph has been provided as fig. C. On the original plate a well-defined pair of dislocations were observed in this area. These have been drawn out schematically in fig. C.

[1] Throughout the present paper hematite has been referred to hexagonal axes, $a = 5{\cdot}035$, $c = 13{\cdot}749$ Å. (Bernal, Dasgupta, and Mackay, 1959).

It is necessary here to state briefly the theory required for the explanation of the fringe formation and its defects. In the diffraction photographs the distribution of scattered radiation is observed as an intensity distribution, which cannot yield any information on the phase of the scattered radiation but only on its amplitude. Accordingly the location of defects cannot be defined. Since in electron-optical resolution, as in an ordinary optical microscope, the scattered electron waves are recombined correctly with regard to both phase and amplitude it is possible to derive an image (with resolution limited only by the objective aperture chosen, and the perfection of the lens system) in which the location of irregularities and defects may be observed directly.

In the present instance the resolution was of the order of 10–15 Å. Study of the present fringe system indicates that the fringes are not all equally spaced. It has also been possible to demonstrate an offset along the length of one particular fringe through a distance corresponding to approximately 10 Å. Apart from indicating the nature of some of the imperfections present, however, the present investigation must be considered to demonstrate conclusively the very high degree of regularity obtaining in the structural state involved. It is obviously not permissible, on the basis of the present experiments alone, to comment in detail on the nature of the inhomogeneity responsible for the well-developed fringe system observed. It is possible only to suggest that a considerable variation in scattering power exists in the direct sequence normal to the fringes, i.e. parallel to c^*, and that this must probably be taken to indicate a corresponding distribution for the heavier iron atoms present in the general structural pattern.

Fringe systems of similar origin have, in the past, been obtained by direct electron-optical resolution of Bragg maxima from suitable crystalline materials (Menter, 1956), and similar resolution experiments illustrating the degree of regularity in antiphase domain patterns have been made by Owaga, D. and H. Watanabe, and Komoda (1958), and by Glossop and Pashley (1959).

The present paper directs attention to the possibility of studying phase transitions of continuous character (i.e. those where a high degree of structural continuity is maintained) by direct electron-optical means, since in this case the nature and distribution of defects may be directly appreciated.

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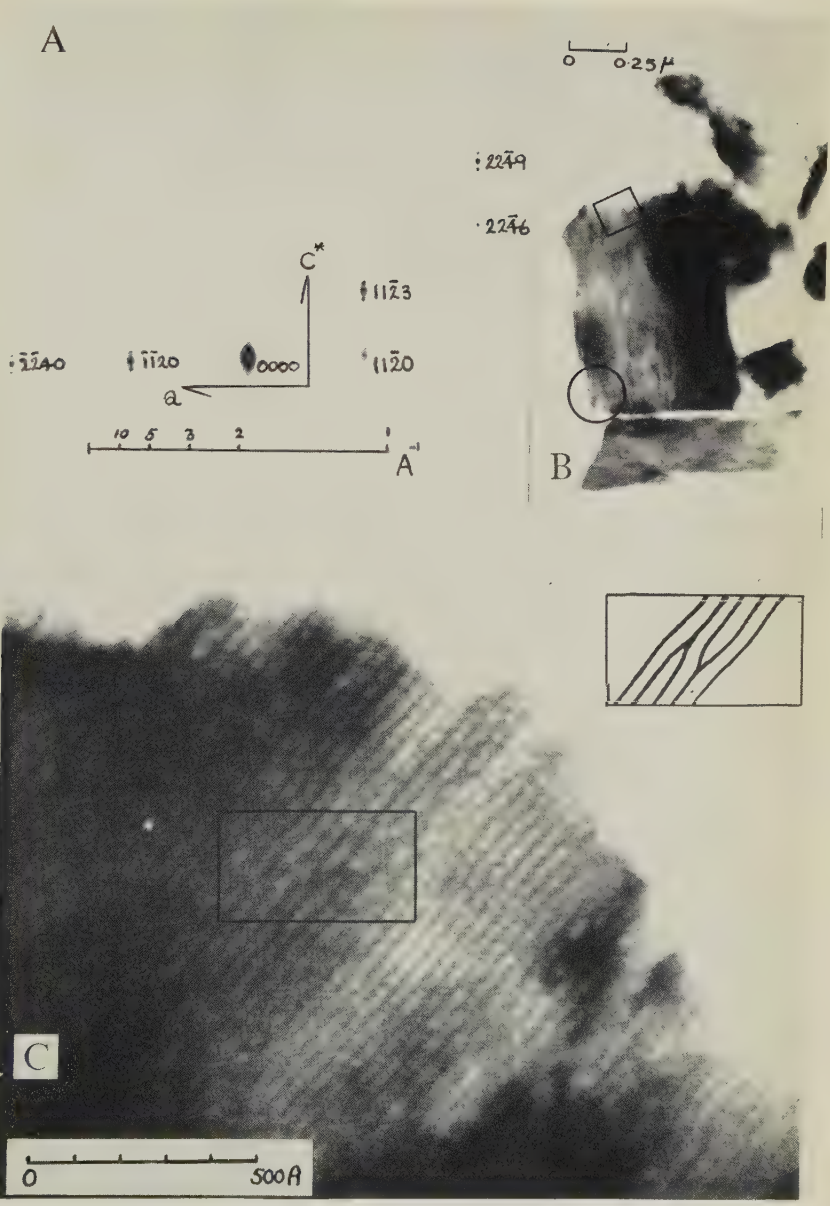

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Explanation of Plate XX.

Fig. A. Indexed electron-diffraction photograph of the $(10\bar{1}0)$ flake studied. The selected area involved has been ringed in fig. B.

Fig. B. Electron photomicrograph of the flake studied. Original magnification $\times$ 20 000.

Fig. C. High resolution photomicrograph of the area defined by a square in fig. B. Original magnification $\times$ 80 000. The *c*-direction in the flake lies normal to the length of the fringes and a pair of dislocations in the pattern have been illustrated.

An X-ray examination of an exceptionally well crystallized kaolinite.

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[Read 8 June 1961.]

Summary. Specimens of kaolinite have been found in inclusions in the ferruginous deposits on the Yorkshire Coast, just north of Scalby. The excellent quality of the X-ray powder pattern suggests that this kaolinite is better crystalline than kaolinites from other localities. Although the lattice constants of the Scalby material are considered to be identical with those previously published for kaolinite, minor differences in the intensities of the X-ray reflections do occur and may be due to a less distorted structural layer in the Scalby kaolinite.

THE material described in this paper was collected from inclusions in the ferruginous deposits on the Yorkshire coast, just north of Scalby. X-ray examination by the powder method revealed that the mineral was kaolinite, whilst observation under the microscope indicated that the crystallite size was ~ 30–$50\,\mu$. The unusual sharpness of the kaolinite reflections in the powder pattern and the relatively large particle size suggest that the kaolinite from this locality is probably better crystalline than any previously reported.

Each kaolinite inclusion was flanked by a layer (a few millimetres thick) of dark brown material, the whole being embedded in a reddish-brown matrix. X-ray powder photographs taken with Co-$K\alpha$ radiation showed that the main constituent of the dark brown material was goethite and that of the reddish-brown chalybite; in each material rather coarse quartz was found in small quantities. The powder pattern of the kaolinite contained a few reflections in excess of those reported by Brindley and Robinson (1946) for kaolinites from other localities. These reflections could not be attributed to any impurity and, in fact, could be explained by supposing that reflections previously recorded as single in the kaolinite pattern were actually resolved in the powder pattern of the Scalby material. This might be expected because the Scalby kaolinite is not so finely divided as other kaolinites, and because the

X-ray technique employed by the authors (see next section) is capable of higher resolution than the one used by Brindley and Robinson.

Two subsidiary tests also suggested that the sample of Scalby material examined was indeed free from impurity. In the first, the sample was treated with hot dilute HCl for several hours without any change occurring in the X-ray pattern. In the second, the sample was heated for three to four hours at 575° C. with a complete disappearance of the accepted kaolinite reflections and also of the additional lines found in the powder pattern of the untreated specimen. Although it can be concluded with some certainty that the sample investigated was pure kaolinite, it must be mentioned that samples from some inclusions were found to contain minor amounts of free quartz.

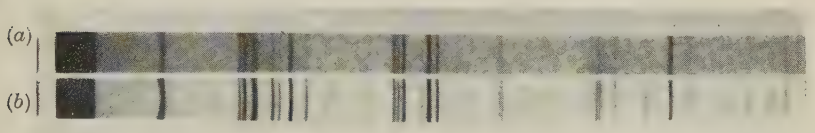

FIG. 1. X-ray powder photographs of (*a*) Cornish kaolinite, (*b*) Scalby kaolinite. Cu-$K\alpha$ radiation, camera diameter 22·9 cm.

The X-ray data. X-ray powder photographs were taken with crystal-monochromatized Cu-$K\alpha$ radiation with a Guinier-type transmission camera of 22·9 cm. effective diameter. The camera was calibrated by mixing aluminium powder with the specimen being X-rayed. Fig. 1*b* shows a typical photograph of the Scalby material, whilst for comparison, fig. 1*a* shows a photograph of a good quality kaolinite from Cornwall. It will be evident that the X-ray reflections from the Scalby kaolinite are sharper than their counterparts from the Cornish specimen. Further, it will be noticed that reflections which appear broad in fig. 1*a* are usually resolved into two reflections in fig. 1*b*. In the case of the Cornish kaolinite, the focusing properties of the X-ray camera did not resolve any reflections that had not previously been resolved by the technique employed by Brindley and Robinson. It thus appears that the greater resolution in the pattern of the Scalby material has arisen mainly because of the larger crystallite size.

The measured spacings and visually estimated intensities of the reflections from the Scalby kaolinite are compared with the corresponding data of Brindley and Robinson in columns B and A, respectively, of table I. Apart from the resolution of certain reflections in the X-ray pattern of the Scalby material, there is little difference between the two

TABLE I. Comparison of powder data of Scalby kaolinite (col. B) with Brindley and Robinson's data (col. A) and with the calculated spacings for a cell having a 5·155, b 8·959, c 7·407 Å., α 91·68°, β 104·87°, γ 89·94° (col. C).

A			B			C
I	$d_{obs.}$	hkl	I	$d_{obs.}$	hkl	$d_{calc.}$
vvs	7·16Å.	001	*vvs*	7·17Å.	001	7·16Å.
m^-	4·46	020	m^-	4·478	020	4·478
m	4·36	$1\bar{1}0$	*s*	4·366	$1\bar{1}0$	4·366
			Tr.	—	110	4·341
m	4·18	$11\bar{1}$	m^+	4·186	$11\bar{1}$	4·180
w^+	4·13	$1\bar{1}\bar{1}$	m^-	4·139	$1\bar{1}\bar{1}$	4·134
wm	3·845	$02\bar{1}$	*m*	3·847	$02\bar{1}$	3·848
w	3·742	021	w^+	3·745	021	3·745
vvs	3·573	002	*vs*	3·579	002	3·578
			Tr.	—	$1\bar{1}1$	3·421
wm	3·372	111	m^-	3·376	111	3·373
w	3·144	$11\bar{2}$	*w*	3·155	$11\bar{2}$	3·155
w	3·097	$1\bar{1}2$	*w*	3·107	$1\bar{1}2$	3·106
w	2·754	022	*w*	2·754	022	2·755
ms	2·558	$1\bar{3}0$ $20\bar{1}$ 130	m^-	2·566	$1\bar{3}0$ $20\bar{1}$	2·568 2·566
			w^+	2·553	130	2·553
wm	2·526	$13\bar{1}$ $1\bar{1}2$	m^-	2·535	$13\bar{1}$	2·534
			vw	2·519	$1\bar{1}2$	2·520
s	2·491	$1\bar{3}1$ 200 112	$m^+(r)$	2·495	$1\bar{3}1$ 200 112	2·504 2·491 2·486
ms	2·379	003	w^+	2·385	003	2·385
vs	2·336	$20\bar{2}$ $13\bar{1}$ $11\bar{3}$	*m*	2·347	$20\bar{2}$	2·346
			m	2·338	$1\bar{3}1$ $11\bar{3}$	2·339
s	2·289	$1\bar{1}3$ 131	*Tr*	—	$1\bar{1}3$	2·307
			m	2·293	131	2·293

Tr, trace reflection.
r, just resolved by eye into two reflections.

sets of data. All the reflections in column B can be accounted for by a unit cell with parameters a 5·155 Å., b 8·959 Å., c 7·407 Å., α $91{\cdot}6_8°$, β $104{\cdot}8_7°$, and γ $89{\cdot}9_4°$; the calculated spacings for this cell are given in column C of table I. The cell agrees, within the limits of experimental error, with that calculated by Newnham (1956) by a least-squares method using Brindley and Robinson's data. Apparently Newnham (private communication) quoted his axial parameters in Å. when in fact they were in kX-units, and this error was carried forward to a later publication by Brindley and Nakahira (1958). The actual parameters of Newnham's cell are a 5·149 Å., b 8·950 Å., c 7·386 Å., α 91·6°, β 104·8°, and γ 89·9°.

Although the lattice of the Scalby kaolinite is indistinguishable from that of other kaolinites, at least as far as the accuracy of the present data is concerned, differences do occur in the relative intensities of some of the reflections. Figs. 2*a* and 2*b* show microphotometer traces of reflec-

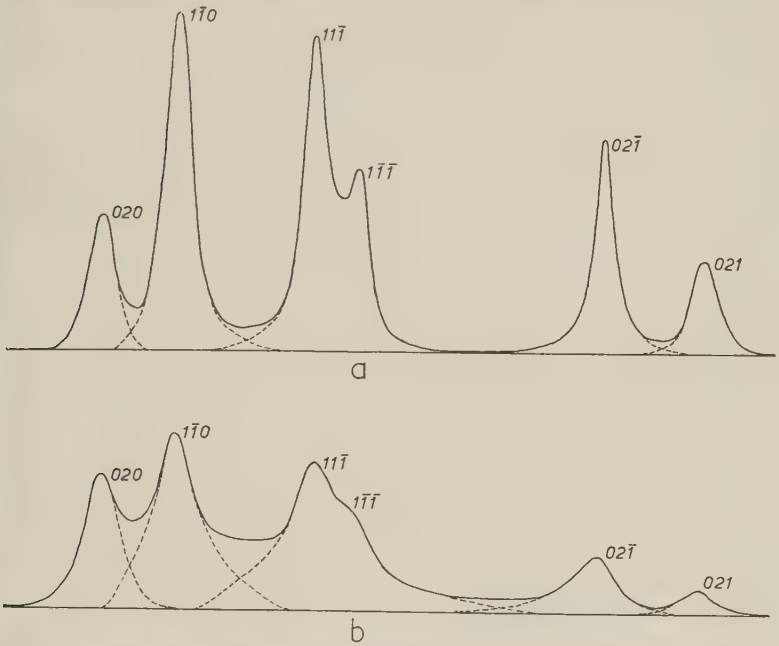

FIG. 2. Microphotometer trace of reflections from (*a*) Scalby kaolinite, (*b*) Cornish kaolinite.

tions, within the range 4·48 Å. to 3·74 Å., from the Scalby and Cornish kaolinites respectively. It is quite noticeable that in the case of the latter, the intensity of the 020 reflection is greater than that of the 02$\bar{1}$, whereas in the Scalby pattern the reverse is true. This is perhaps the most obvious difference, but others do occur. It is not thought that the intensity data have been modified significantly by possible preferred orientation in the particular form of specimen used in the present technique. This is supported by the facts that the relative intensities of the reflections from the Cornish kaolinite agree very well with those obtained for a number of kaolinites by Brindley and Nakahira using a diffractometer method, and that X-ray photographs of the Scalby material taken with a 19-cm. diameter Debye-Scherrer camera showed the same pattern of intensities as was obtained using the Guinier-type camera.

Discussion of the kaolin-type layer in the Scalby mineral. According to the conclusions reached by Brindley and Nakahira, the intensity differences in the X-ray data of the Scalby and Cornish kaolinites could possibly be due to a smaller degree of distortion in the structural layer of the Scalby mineral. Brindley and Nakahira explained the triclinic nature of kaolinite by considering the kaolin layer to have the same degree of distortion, regarding rotations of Si–O tetrahedra and Al–O(OH) octahedra, as was found to occur in dickite by Newnham and Brindley (1956). They also found that a dickite-type layer led to an improvement on earlier calculations in the agreement between the calculated and observed intensities in the kaolinite powder pattern, and that rather better agreement was possible if one considered an even more distorted layer than that found in dickite.

Calculated relative intensities (taken from Brindley and Nakahira's paper) of a series of kaolinite reflections are given in columns A and C of table II; the data in column A are for a structure based on the dickite-

Table II. Comparison of calculated and observed intensities for kaolinites from Scalby and Cornwall.

	A.	B.	C.	D.
hkl.	$I_{calc.}$	$I_{obs.}$	$I_{calc.}$	$I_{obs.}$
020	100	10	100	10
$1\bar{1}0$	273	27	215	22
110	9		2	
$11\bar{1}$	292	36	205	25
$1\bar{1}\bar{1}$	133		96	
$02\bar{1}$	219	16	104	6
021	67	7	49	3

Columns A and C give calculated values of $F^2\phi(\theta)$ for structures based on a dickite-type layer and on a more distorted layer respectively, ϕ (θ) being the polarization and Lorentz factor.

Columns B and D give the observed intensities for the Scalby and Cornish kaolinites, respectively.

Data in each column are scaled so that the intensity of the 020 reflection is taken as standard.

type layer and those in column C for a structure with a more distorted layer. It will be seen that the observed intensities for the Scalby kaolinite (column B, table II) are, in general, in better agreement with the calculated values in column A, whilst those for the Cornish kaolinite (column D, table II) are more consistent with the data in column C. This suggests, though very tentatively in view of the small number of reflections considered, that the kaolin layer in the Scalby material is less

distorted than that in the Cornish kaolinite, or apparently in kaolinites that have been investigated previously.

Because of the inadequate resolving power of the microphotometer used in this work, it did not seem worthwhile to attempt to measure quantitatively the intensities of the other reflections considered by Brindley and Nakahira. Nevertheless these are explained qualitatively by the above conclusions.

Acknowledgement. The authors are indebted to Professor G. W. Brindley for much helpful information.

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SHORT COMMUNICATIONS

A note on manganophyllite from Goldongri, India.

In 1957, during general reconnaissances of the Goldongri manganese mine, Panchmahal district, Gujrat State, India, a few specimens were collected that on examination revealed an interesting manganiferous mica. The occurrence of manganophyllite from this area (Jothvad) was reported by the late Sir Lewis Fermor.[1] As far as is known to the author no work has been undertaken on this manganophyllite since Fermor's time.

The mica occurs as small flakes and sometimes forms aggregations of thin scales in quartzite and in a winchite–blanfordite rock. It varies from brown to reddish brown with a bronzy lustre. Under the microscope it is observed that the mica is distributed throughout the whole section but especially along the cleavages of the amphibole and pyroxene. The optical characters of the mica are: $\beta = 1{\cdot}59$–$1{\cdot}60$; $2V_\alpha = 20°$–$32°$; the optic axial plane is parallel to {010}; pleochroism, α pink with an orange tinge, β and γ yellowish brown to dark brown, absorption $\gamma = \beta > \alpha$. The optical characters together with the presence of manganese indicate that the mica belongs to the manganophyllite group of Fermor. There is, however, great variation in pleochroism, in a few cases some lamellae interlaminated with the manganophyllite lamellae show pleochroism similar to that of biotite; owing to this interlamination of manganophyllite and biotite the determination of its pleochroic scheme is a matter of difficulty. Other associated minerals are: predominantly winchite, blanfordite, biotite, quartz, microcline, plagioclase, and manganese ore of braunitic nature.

The optical characters of the mica are very similar to those of manganophyllite described by Rao[2] from Tirodi mine, Madhya Pradesh, India. In pleochroism it differs slightly from the manganophyllite described by Fermor from the present area. The cause of variation in pleochroism is not clearly understood. Fermor reported the presence of both uniaxial and biaxial manganophyllites.

Optic axial angles for reported Indian manganophyllite range from 0 to 40° (table I). The cause of this variation is not known: Rao[2] opines that it may be due to the effect of manganese or iron on the optical characters of the mineral. In this connexion it will not be out of place to mention the suggestion made by Fermor (p. 196): 'Some of these micas are uniaxial and some of them biaxial, sometimes strongly so. It is

possible that some of them may be manganophyllite, but it is almost certain that there are two or more species amongst them.'

TABLE I. Optic axial angles of Indian manganophyllites.

Reported by	Locality	$2V_a$
Fermor (1909)	Jothvad (Goldongri)	Some uniaxial and some biaxial
Fermor (1909)	Sitapathur	Uniaxial
Fermor (1909)	Junawani	Biaxial
Kilpady & Dave (1954)	Sitapathur	10°
Bilgrami (1952)	Sitasaongi	18°–22°
Kilpady & Dave (1954)	Sitasaongi	Between 35° and 36°
Raghava Rao (1955)	Tirodi	23°–40°
Basu (1958)	Ramtek-Chorbaoli	10°
Present author	Goldongri (Jothvad)	20°–32°

From the available data on manganophyllites it is apparent that there is a continuous variation in the optical axial angle, which probably indicates that there is a continuous variation in the composition of the mica and it seems difficult to demarcate it into any recognizable species.

Bilgrami,[3] Kilpady,[4] Basu,[5] and Rao[2] suggested that the manganophyllite has developed from the amphibole, with which it is constantly associated. In the present case, however, it does not seem to have developed from the amphibole; under the microscope it is observed that manganophyllite occurs both in the winchite–blanfordite portion and also in the quartz–feldspar portion of the rock and therefore its origin cannot be stated with certainty, but considering the fact that the granite and pegmatite intrusives are found in close proximity it is not unlikely that it may have had an igneous origin though the necessary manganese was probably absorbed from the adjacent manganese ore body.

The author thanks M/S Sundar Minerals (P) Ltd., for permitting him to visit the mine and for facilities offered during the work, and also the Ministry of Scientific Research, Government of India, for the award of a research scholarship, which enabled him to carry out this work.

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The Bamenda 'meteorite'.

Fragments of the mass from Bamenda, Cameroons, Nigeria, described by M. D. W. Jeffreys (1955)[1] as a meteorite, were recently received at the British Museum (Natural History); it was evident that the material was a fairly pure maghemite, and analysis confirmed this: Fe_2O_3 90·7, FeO 4·6, TiO_2 0·1%, Cu 80 p.p.m., Ni absent, H_2O present and amounting to 4·6% (by difference). The name Bamenda should be struck out from the list of meteorites.

British Museum (Natural History),
London, S.W. 7

M. H. Hey

[1] M. D. W. Jeffreys, *Man*, London, 1955, vol. 55, p. 167 [M.A. **13**–81].

Humite and chondrodite in a Lewisian crystalline limestone from South Harris, Outer Hebrides.

The presence of chondrodite in the Lewisian has only previously been recorded from the crystalline marbles of Glenelg (Read and Double).[1] During a recent re-examination of the South Harris metamorphosed limestones an isolated occurrence of humite and chondrodite has been found at Bay Steinigie near the Finsbay to Borve paragneiss belt. The Bay Steinigie paragneisses, which include diopside-amphibolites, quartzose biotite-schists and occasional lenses of forsterite- and diopside-marbles, form a long narrow xenolith within the metamorphosed intermediate (tonalitic) rocks of the South Harris igneous complex.

In one of these lenses a very small amount of humite and chondrodite is present. The rock consists mainly of calcite, dolomite, and rounded, partly serpentinized grains of forsterite, with which small grains of humite and chondrodite are sometimes associated. Other minerals present include phlogopite, pargasite, small laths of brucite, and minor amounts of sphene, graphite, and opaque iron oxides.

The humite and chondrodite typically occur in small yellow rounded grains (about 0·30 mm. across) and are usually found around the margins of the partly serpentinized forsterite crystals. They are often associated with brucite. The few grains in which cleavage is present may either show extinction parallel to this (001) parting (humite) or inclined at an angle of about 25° (chondrodite). The humite and chondrodite grains are highly birefringent and have pleochroism: α golden yellow, β and

γ pale yellow, $\alpha > \beta = \gamma$. One small elongated humite grain showing (001) parting and parallel extinction has a positive optic axial angle of about 68°.

In an extensive collection of specimens from the crystalline limestone lenses of the two paragneiss belts of South Harris, humite and chondrodite have only been found at this one locality in Bay Steinigie.

Geological Survey and Museum,
Exhibition Road,
South Kensington,
London, S.W. 7.

R. Dearnley

[1] Min. Mag., 1935, vol. 24, p. 84.

BOOK REVIEWS

Korzhinsky (D. S.). *Physicochemical basis of the analysis of the paragenesis of minerals.* Consultants Bureau, Inc. (New York), and Chapman & Hall (London), 1959. 142 pp., 86 text-figs. Price $7.50 or 63*s*.

Korzhinsky is concerned with elucidation of the factors that have led to the simultaneous formation of mineral assemblages, especially in metamorphic and metasomatic rocks. Relict and retrogressive minerals are therefore excluded from his analysis. In his view, processes of mineral formation represent reactions between two systems, one comprising the minerals in a rock and the other a pore fluid. Components of these systems are termed inert and mobile, respectively. He suggests that the circulation of mobile components in a chemically active pore fluid is essential for the occurrence of metamorphism. Korzhinsky sets up a thermodynamic model in which the chemical potentials of inert components are dependent upon their fixed masses and those of the mobile components are controlled by external conditions. The mobile components are then 'perfectly mobile' and their masses in a rock at any equilibrium stage depend only upon the content of inert components and the external conditions. The factors of equilibrium controlling mineral parageneses are thus the masses of the inert components (extensive parameters) and the intensive parameters pressure, temperature, and the chemical potentials of perfectly mobile components. The purpose of the book is to study the dependence of mineral parageneses on

these known factors of equilibrium. This Korzhinsky regards as a necessary basis for the reverse process of discovering the factors of equilibrium during natural processes from the study of mineral parageneses in rocks. Since natural systems only approximate to true thermodynamic equilibrium the 'relative mobilities' of mobile components have to be considered.

The approach can be illustrated by the content of successive chapters. I, Thermodynamic basis of the analysis of paragenesis of minerals, 27 pp. Gibbs's phase rule is applied to open systems with perfectly mobile components. II, Methods of representing chemical composition, 27 pp. Methods of projective geometry are emphasized. III, Relation between chemical and mineralogical composition under constant external conditions, 19 pp. Triangular composition paragenesis diagrams are used to illustrate mineral parageneses, and multicomponent systems are resolved into partial systems containing only three inert components. Consideration of metamorphic, metasomatic, and igneous processes leads to a discussion of mineral facies. IV, Dependence of mineralogical composition on external conditions, 56 pp. Changes in external conditions (intensive parameters) cause changes in mineral parageneses and mineral composition that are illustrated in composition–paragenesis and chemical potential diagrams. The latter permit illustration of all possible parageneses in a multicomponent system on one plane. The chemical potentials of H_2O and CO_2 are frequently used as co-ordinates for metamorphic reactions, and those of K_2O and Na_2O for igneous reactions. V, General remarks on the application of paragenetic analysis of minerals in geology, 3 pp. Numerous observations of mineral parageneses over large areas and covering the entire range of composition are required before the data can be successfully analysed with the consequent development of physico-chemical theories of mineral formation.

The translation is good and there are few misprints; the addition of an index would have been most welcome. Although much of the text is theoretical there are many geological examples selected from Korzhinsky's own extensive field and petrographic studies. The reviewer is unable to comment on the rigour of the thermodynamics but he is convinced that the paragenetic diagrams will find wide application when the methods are more generally known. A useful sequel would be an enlargement of chapter V. The book can be recommended to all geologists interested in processes of mineral formation.

P. J. W.

GINZBURG (I. I.). *Principles of Geochemical Prospecting.* Translated from the original Russian by V. P. Sokoloff. Pergamon Press, 1960. (70*s*.)

Geochemical prospecting, as defined by Hawkes (1957, p. 226) '. . . includes any method of mineral exploration based on systematic measurement of one or more chemical properties of a naturally occurring material. The chemical property measured is most commonly the trace content of some element or group of elements: the naturally occurring material may be rock, soil, gossan, glacial debris, vegetation, stream sediment or water. The purpose of the measurements is the discovery of a geochemical "anomaly" or area where the chemical pattern indicates the presence of ore in the vicinity.'

Ginzburg's monograph is a critical summary of applied geochemical investigations that have been carried out during the past 25 years by Soviet scientists in areas differing widely in geologic and climatic character, and is based on the data in the papers presented at the All-Union Conference on Geochemical Methods of Prospecting for Ore Deposits, which was held in Moscow in 1956.

The book is to be recommended to all who are concerned with the search for ore or with geochemistry in the widest sense of the word. Apart from Hawkes's small, but excellent, 'Principles of Geochemical Prospecting' (1957), it is, as yet, the only text-book on the subject, and these two works are, in many respects, complementary to each other. In addition, it contains a vast amount of material not hitherto readily available to those who do not read Russian. Clearly, the work carried out in this subject in the U.S.S.R. cannot be ignored when it is remembered that for several years 6 to 7 million soil samples have been analysed annually there to facilitate the search for ore, and that since 1957 every area that is geologically surveyed must also be subjected to geochemical investigation.

The present English translation was made by V. P. Sokoloff, a Russian-born scientist, who has had considerable experience in applied geochemistry in the United States and elsewhere. He has lavished care on this work, and his critical approach to the task is indicated by his illuminating footnotes and by the glossary of Russian soil terms that he has compiled.

Although investigations carried out by non-Russian workers have not been ignored, and 43 of the 313 references in the text relate to such work, yet incomplete knowledge of the contribution made by these has led to

statements that are *not* correct. Thus, on p. xvii it is stated that 'sizeable deposits of tin and tungsten were discovered in England during the last fifteen years, by geochemical surveys, in Cornwall, Devonshire, and Wales . . .'. Apart from the obvious geographical error, neither the Swedish Company nor others who have employed geochemical prospecting in these areas have found a mineral deposit of economic importance. On p. 8 Ginzburg remarks that chemical methods of analysis for small quantities of arsenic and antimony that would be suitable for applied geochemical investigations 'are either inadequate or entirely undeveloped'. In point of fact perfectly adequate methods have been known in Britain for several years. It is, however, clear that the Russians have generally employed spectrographic methods of analysis rather than colorimetric ones. Ginzburg advocates the greater use of the latter and he includes details of a few of these methods in an appendix.

The 72 diagrams considerably enhance the value of the work and they have gained much by being redrawn from the original, but the fundamental construction of some is such that they cannot be readily understood, and, unfortunately, many of the maps lack scales and cardinal points.

Despite the above criticisms the monograph is a major contribution to applied geochemical literature and the extent to which the subject is covered may be judged by the following chapter, &c., headings: Introduction (essentially a historical survey). Methods of analytical investigations. Geochemical tracers (indicators). Accumulation of metals in igneous and metamorphic rocks. Accumulation of metals in sedimentary rock. Prospecting in bedrocks. Prospecting for deposits without surface outcrops. Accumulation of metals in unconsolidated overburden (dispersion halos). Prospecting at the surface of the overlying mantle. Migration of metals in waters. Hydro-geochemical prospecting for metals and characteristics of different water types associated with ore deposits. The bio-geochemical method of prospecting. General conclusions in reference to geochemical survey. Appendix. Rapid methods of determining Pb, Cu, W, Mo, and Ag in the field. Glossary. Translator's notes on soil terms used in this book.

K. F. G. Hosking

Kasper (J. S.) & Lonsdale (K.), Editors. *International Tables for X-ray Crystallography, Vol. 2: Mathematical Tables.* Birmingham, England (Kynoch Press for the International Union of Crystallography), 1959, xviii+444 pp., 39 figs. Price 115*s.*; there is a conditional reduced price for individuals on application to the Printers.

The first volume of this new series was published in 1952. This second volume covers a larger and more varied field and is the work of a number of authors. The main sections are: Introduction, Fundamental Mathematics, Crystal Geometry, Diffraction Geometry, Physics of Diffraction Methods, Fourier Synthesis and Structure Factors, Special Topics, Miscellaneous Exponential and Trigonometrical Tables. The work is completed by a Dictionary of Crystallographic Terms in English, French, German, Russian, and Spanish. Items of special interest to the mineralogist are the sections on twinning and on the interpretation of single-crystal and powder photographs, but the volume contains much that is of use to all those engaged in different branches of crystallographic studies. The Editors have done a splendid job, and the whole production is of the high standard that would be expected from the Printers and from the International Union of Crystallography.

C. H. Kelsey

Williams (Howel). *Volcanic History of the Guatemalan Highlands.* University of California Publications in Geological Sciences, vol. 38, No. 1, pp. 1–86, pls. 1–13, 8 figs. (including 2 folding maps).

This account is based on field reconnaissances totalling four months under a Guggenheim fellowship. Apart from a succinct preliminary sketch of the basement geology, it is entirely devoted to detailed descriptions of the outcrops of the volcanic rocks, the physiography, and the history as deduced from these. As the only up-to-date general account of this area it is most valuable, but it contains no detailed petrographic descriptions and the author's use of rock names is vague; one suspects from his descriptions that most of the rocks called 'andesite' and 'dacite' would have been named 'rhyodacite' had analyses been available. The author throughout commendably uses the term 'glowing-avalanche' instead of 'nuée ardente'. A feature of his observations is the extent to which he has recognized the prevalence of this type of deposit especially in the first, Tertiary, 'fissure-eruption' phase of vulcanicity. He recognizes, however, that most of the vast pumice deposits of the second, Quaternary phase are of secondary, penecontemporaneous, torrential (laharic) origin. Much attention is devoted to the geomorphological history of the area, especially the origin of its three main lakes: Atitlan (as a cauldron subsidence), Ayarza (as paired calderas), and Amatitlan (as a volcanotectonic basin).

S. E. Ellis

WADE (F. A.) and MATTOX (R. B.). *Elements of Crystallography and Mineralogy.* New York (Harper & Brothers—Harper's Geoscience Series), 1960, xiv + 332 pp., 18 tables, 327 text-figs. Price $7.50.

Pleasing to read and clear in layout and presentation, this elementary textbook is arranged in conventional manner and although modern in outlook has no pretensions to startling innovation. The authors state in their preface that they intend the book to be used in conjunction with a manual of determinative mineralogy (unspecified) and that they deliberately omit optical crystallography—so thoroughly, in fact, that the chapter on physical properties contains not even a passing mention of refractive index.

There are ten chapters, followed by an appendix (4 pp.) on stereographic projection, a mineral index (9 pp.), and a subject index (6 pp.): Introduction (4 pp.); A review of the concept of matter (10 pp.); The crystalline state (22 pp.); The classification of crystals (79 pp.); Crystal chemistry (20 pp.); Physical properties (12 pp.); Chemical properties (26 pp.); Descriptive mineralogy (99 pp.); Economic mineralogy (18 pp.); Genetic mineralogy and mineral associations (22 pp.). The chapter on the crystalline state is brief but adequate, and contains the elements of symmetrical and metrical crystallography. The chapter on the classification of crystals is concerned mainly with the 32 classes, with a short final section on twinning, habit, and types of aggregation; the treatment of the 32 classes is far too detailed for the intended readership, and the nomenclature is archaic—do we need to consider a 'positive left-handed rhombohedron of the third order' (p. 82) or a 'negative fourth-order prism' (p. 94)? The distinction as one of orders serves no useful purpose (F. C. Phillips, Introduction to Crystallography, 2nd edn, p. 65) and in any case the terms are not put to use in the descriptive section. The chapters on physical and chemical properties need little comment other than that made earlier, and that X-ray diffraction is rather oddly considered in the chemical section. The descriptive mineralogy seems to receive adequate treatment—the reviewer did not read every description, but noted that the commonest colour of phenakite (colourless) was omitted in favour of yellow, pink, or brown (p. 238), and that five of the six figures (on silicon–oxygen skeletons) had already appeared in identical form in the crystal chemistry chapter. The economic and genetic chapters are limited but good. The most damaging criticism of the book is that too little care has been taken in the correction of errors, and that several of the crystal drawings are horrible (especially figs. 4.12, 4.15, 4.32, and 4.33) in that parallelism of edges in a zone is frequently ignored

and that the cubic axial cross has the y and z axes at 90° in projection; figs. 4.148 and 4.149 (right- and left-handed quartz) have their captions interchanged. Wrong formulae appear, such as (Fe, Mn)WO_4 for wulfenite (table 4.2), $MgCl_2.6H_2O.8B_2O_3$ for boracite (p. 303), and $Ca_8B_{18}Cl_4.4H_2O$ for parahilgardite (table 4.8 and p. 322). Minor mis-spellings are not uncommon (olivene for olivine on p. 142, lazullite for lazulite, triphyllite for triphylite on p. 222, and Moh's for Mohs' on pp. 141 and 142, &c.), and major examples include lithiophyllite for lithiophilite (pp. 223 and 321) and desclosite for descloizite (p. 319). Astrakanite (spelled astraconite and astracanite), schönite (spelled schonite), and penta salt ought to give way to blödite, picromerite, and görgeyite respectively.

P. G. E.

Swineford (A.), editor. *Clays and Clay Minerals.* Volume 8. Proceedings of the Eighth National Conference on Clays and Clay Minerals. London (Pergamon Press Ltd.), 1960, ix + 292 pp., 128 text-figs. and pls., 51 tables. Price 63*s*.

Like its predecessors [M.A. **13**–313; **13**–314; **13**–428; Min. Mag. **32**–258] this volume contains, either in full or in abstract, the papers presented at a National Conference on Clays and Clay Minerals (the 8th) sponsored by the Committee on Clay Minerals of the National Academy of Sciences—National Research Council and the University of Oklahoma, held at Norman, Oklahoma, on 11–14 October 1959. The first paper gives a well-documented account of the field excursion associated with this Conference. Of the other 21 papers and abstracts, 6 deal with clay–water or clay–solution aspects and 4 with geological or prospecting matters, the remainder covering such subjects as ion exchange, clay–organic complexes, alteration and synthesis, and diagnostic techniques. It would be impossible in a short review such as this to deal with each paper individually and it would be invidious to make a selection. Generally, however, it may be remarked that the quality of the papers is rather higher than in some of the previous volumes, and there are very few where one feels the authors might have developed their subjects more intensively. The general impression one receives is that the advances made in clay mineralogy over the past few years have been such that quantitative interpretation of phenomena, particularly in clay–water systems, is now possible by rigid mathematical and physical processes and that clay mineralogy is consequently rapidly developing from a somewhat speculative into an exact science. The amount of quantitative data presented in this volume is particularly striking. The

abstracts given at the commencement of each paper are concise and indicate very clearly the scope of the paper and the conclusions reached. The standard of presentation and publishing is excellent and the book is provided with a brief but adequate index. In the papers generally, there is more reference to work done outside America than has often been the case in previous volumes. The editor and printers are to be congratulated on the high standard attained and on providing a volume which will not only be essential to clay mineralogists but also to those interested in rheological and geological problems and to several branches of industry.

R. C. Mackenzie

Russell (R. D.) and Farquhar (R. M.). *Lead isotopes in geology.* New York and London (Interscience Publishers), 1960, 243 pp. Price $9.00 or 65*s*.

This monograph deals principally with the use of common lead in geochronology. As it includes a series of tables for deriving ages it is necessary to examine with some care the parameters selected for the basic equations.

Mathematical models for dating ordinary lead are of two kinds, depending on whether they postulate a uniform distribution of lead, uranium, and thorium in the source material or allow variation from place to place. In both cases the lead has grown from primeval lead by the addition of radiogenic isotopes from uranium and thorium. According to the first hypothesis, leads formed at the same time have the same isotopic composition. On the second, the composition depends on the proportions of primeval lead, uranium, and thorium in the source material as well as on age, but since ^{206}Pb and ^{207}Pb are decay products of a single element, there is a constant relationship between the ratios 206/204 and 207/204 in all leads of the same age; this is the basis of the Holmes–Houtermans method of dating. The age equations on both hypotheses incorporate the decay constants and the present-day $^{238}U/^{235}U$ ratio, all of which are tolerably well known, and make specific assumptions for the compositions of primeval and modern lead. Put in another way, the compositions of primeval lead (known from uranium- and thorium-free meteorites) and of a single ordinary lead of known age are sufficient to establish the equations for either type of model and, incidentally, to give the age of the earth.

The particular Holmes–Houtermans model in this book uses values

for the constants and for primeval lead differing slightly from those in Houtermans' original equation, but the parameters have been adjusted to give closely agreeing ages. Houtermans specified his modern lead so that a number of Post-Cambrian leads had ages in keeping with their geological relationships. No doubt this empirical approach is the right one, but the actual values adopted for modern lead should be reviewed periodically in the light of refinements to the time scale and of new lead analyses.

For the homogeneous-source hypothesis, which is open to criticism on general grounds as being at variance with the well-established concept of geochemical provinces, the reader is offered a choice of two models. The first one, designated the Russell–Farquhar–Cumming method, can be discarded at once as being inconsistent with the meteoritic ratios for primeval lead. It is claimed that the second, designated the Russell–Stanton–Farquhar method, gives the age of lead in stratigraphically conformable deposits rather accurately, if Stanton's theory that such lead is derived directly from the upper mantle is accepted. Two ages are obtained from the ratios 206/207 and 208/204 respectively. The growth curve for the first is defined as starting at the meteoritic ratios for primeval lead, passing through the measured ratios of lead from a particular conformable deposit, and finishing on a zero isochron derived from Patterson's analyses of the total-leads in three uraniferous meteorites. This zero isochron differs considerably from that adopted for the Holmes–Houtermans model—a lead sample of zero age on the Russell–Stanton–Farquhar model has a Holmes–Houtermans age of 204 million years. Being based on lead ratios for which Patterson claimed only 2% accuracy, the R-S-F zero isochron is likely to be much the less accurate. Its unreliability for this purpose is indicated by the spread of the Holmes–Houtermans ages calculated for the total-leads from the three meteorites: +301, −32 and −454 m.y. The other R-S-F growth curve, that of the 208/204 ratio, has been defined so as to give the best fit for galenas from eight conformable deposits, but it is inconsistent with the 208/204 ratio for primeval lead and with the age of the earth implied by the 206/207 equation.

Leads that have been enriched in radiogenic isotopes during migration of the ore fluids through crustal rocks are 'anomalous' in that the simple models discussed above are inapplicable. Such leads, however, can provide valuable information on the history of the ore deposits in which they occur, as is well brought out in case studies of Sudbury and of Thunder Bay, Ontario, and of Broken Hill, N.S.W. A potential

hazard in dating common leads is that anomalous samples may not be recognized as such if the added component is small, with the result that erroneous ages are derived. The danger is greatest when reliance is placed on single specimens, a practice the authors rightly condemn. A representative suite should be collected and the isotopic analyses considered in the light of the known geological environment of the samples.

Recent developments in dating uranium and thorium minerals are discussed in outline only, as the authors consider this aspect of their subject has been adequately covered elsewhere. This is unfortunate, since the validity of ages that 'provide the standards against which all other methods of age determination are compared' is important enough to warrant detailed treatment in any book on lead isotopes in geology.

A series of appendices, filling 121 pages or half the book, lists several hundred isotopic analyses of common lead. About half these are Toronto analyses, both published and unpublished; the remainder are published analyses from other laboratories. The analyses are grouped geographically, but geological associations are seldom given owing to lack of data.

Attention should be drawn to misprints that might mislead the reader: p. 50, l. 2, V should be '0.0800', not '11.0(2)'; p. 65, eq. (5.4), the exponential term should be in the denominator; p. 74, l. 11, '0.128' should read '1.28'; p. 85, l. 11, 'grain' should read 'gramme'; pp. 158, 159, the geographical heading is incorrect. Errors have been noted as follows in the tables for deriving ages from the proposed equations: Table 4.1 −372 (should be −327), 1(61), 3701(3710); Table 4.2 230(232); Table 4.5 3081(3181); Table 4.6 2370(2367), 386(406), 329(349); Table 8.1 890(887).

The book is in the nature of an interim report on a rapidly developing subject. It was perhaps too hurriedly put together and was certainly inadequately revised, but it offers a stimulating discussion of much scattered literature and no one interested in geochronology or in lead ores would wish to be without it.

J. E. T. Horne